PR
00
86

THE JOURNAL OF SOIL SCIENCE

UNIVERSITY OF MICHIGAN
NOV 3 ~ 1955
SCIENCE LIBRARY

VOLUME 6 NUMBER 1
JANUARY 1955

OXFORD
AT THE CLARENDON PRESS

THE
JOURNAL OF SOIL SCIENCE

Editor: G. V. JACKS, M.A.

COMMONWEALTH BUREAU OF SOIL SCIENCE
ROTHAMSTED EXPERIMENTAL STATION, HARPENDEN, HERTS.

NOTICE TO CONTRIBUTORS

The editorial policy is to accept papers on original research or reviews dealing with the science of the soil in its widest aspect providing they are of a high scientific standard, and research workers whether within or without the British Commonwealth are invited to offer papers, in English, to the Editor.

Contributors are asked to submit their papers in typescript in a final and finished state, with double-line spacing and ample margins. An allowance of sixteen shillings per sheet of sixteen pages is made for alterations in the proof, contributors being responsible for any excess. Illustrations should if possible be drawn on Bristol board in Indian ink with lettering inserted lightly in pencil. Drawings should be rather larger than the size of the printed block, and their order and approximate position in the text should be marked. Tabular matter should be kept to a minimum. References should give in the text, author and date in brackets, and at the end of the paper appear in alphabetical order in the form: author's name (surname first), date, title of paper, name of journal, volume number, page numbers.

Contributors are asked to write on their papers the address to which proofs are to be sent, and to state, at the time when they return corrected proofs to the editor, if they wish to buy offprints in addition to the twenty-five copies which are allowed free of charge.

THE JOURNAL OF SOIL SCIENCE is published twice yearly and two numbers constitute a volume. Price, per number, 17*s.* net, by post 17*s.* 6*d.*; per volume (pre-paid) 30*s.*, post free. Orders may be sent to the Publisher or any bookseller.

Members of the British Society of Soil Science receive the Journal free of charge. Application for membership (annual subscription £1. 1*s.* 0*d.*) should be addressed to the Treasurer, Dr. W. E. Chambers, Royal Agricultural College, Cirencester, Glos.

Correspondence on the subject-matter of the Journal should be addressed to G. V. Jacks, Esq., Commonwealth Bureau of Soil Science, Rothamsted Experimental Station, Harpenden, Herts., and all other correspondence to the Publisher:

GEOFFREY CUMBERLEGE
OXFORD UNIVERSITY PRESS
AMEN HOUSE, LONDON, E.C.4

ERRATUM

VOL. 3, *p.* 254. The legend of FIG. 2 should read:

FIG. 2. DTA standard kaolinite: Calcined kaolin mixtures. Endothermic reaction (595–625°C.) Curve 1. Total weight of active material – 430 mg. full crucible. Curve 2. Total weight of active material – 200 mg. 'sandwich'-packed. Curve 3. Reproduced from FIG. 1.

STUDIES IN PEDOGENESIS IN NEW SOUTH WALES

VII. THE 'GILGAI' SOILS

E. G. HALLSWORTH,* GWEN K. ROBERTSON,† AND F. R. GIBBONS

(*Agricultural Chemistry Laboratories, University of Sydney*)

A FEATURE of some interest that is widely developed on certain soils of New South Wales and elsewhere throughout Australia is the occurrence of small-scale undulations of the surface of the land, the alternate hummocks and hollows of which show some degree of regularity. These have been called variously 'gilgai', 'crab-hole', 'melon-hole', 'devil-devil', and 'Bay of Biscay country', and were first described in soil literature by Prescott (1931) and subsequently by other writers (e.g. Leeper, Nicholls, and Wadham, 1936; Howard, 1939).

Considerable differences in magnitude and form of the undulations occur, and since the different names are not applied consistently to any one form, the term 'gilgai'‡ has been used here collectively for all the different manifestations of what appears to be essentially the same process.

Two further points add interest to the problem. The first is that after the gilgaies have been levelled by grading and cultivation, they will generally develop again in from 2 to 11 years. This indicates that whatever was responsible for their formation is still generally operative at present. The second point is that patches of country carrying no gilgaies are found scattered across areas of otherwise similar country covered with them, and it is common to find all gradations from well-developed gilgai to a self-mulching soil occurring across an area.

Not only the forms, but also the individual features are varied. The lower parts of the surface may be flat or smoothly concave and may carry towards the centre one or more holes where the surface soil has broken and fallen in. It is from these holes that the terms 'crab-hole' or 'melon-hole' originated. The depressed or lower portion of the soil surface has been termed the 'shelf', whilst the hummock or raised part has been termed the 'puff', the latter from the well-structured and 'puffy' surface soil that characterized the hummocks of the gilgaies of the northern Riverina originally described by H. N. England (Prescott, 1931). This puffiness of the hummock is not universal, and is lacking not only from the 'melon-holes' of the Richmond river, but also from the 'Stony Gilgaies' of the West Darling.

Environmental conditions are very diverse, and in New South Wales gilgaies are found under rainfall varying from 60 in. in the Tweed valley to less than 6 in. west of Tibooburra, but always associated with high summer temperatures. The parent materials are basalt, calcareous fine-

* Now of the University of Nottingham, England.

† Most of the results in this paper have been taken from the thesis presented by G. K. Robertson for the degree of M.Sc. of the University of Sydney.

‡ Taken from an aboriginal word for a small water-hole.

Journal of Soil Science, Vol. 6, No. 1, 1955.

5113.6.1

grained sedimentary rocks, or fine-grained alluvial deposits. The associated vegetation is equally diverse and varies from wet sclerophyll forest through sparse savannah woodland to grassland and salt-bush in the far west.

The gilgai formations occur as variants of a number of distinct soil groups, and have been observed on meadow, chernozemic, sierozem, and stony downs (stony-desert) soils.

Classification

All the gilgaies observed to date in New South Wales can be classified in six main types, although many intermediate forms exist. These have been termed normal gilgai, lattice gilgai, wavy gilgai, tank gilgai, stony gilgai, and melon-hole gilgai respectively.

1. *Normal or round gilgai.* These are the most widespread. The puffs and shelves are sub-circular in outline, usually equally well developed, and show no particular orientation. The shelves are mostly concave and not uncommonly without a sink-hole. The degree of development varies from cases showing no perceptible vertical interval, and where the feature may be detected only by the change in colour of the surface soil and the presence of carbonate concretions on the puff, to cases where the vertical interval is 8 ft. and the wave-length up to 40 ft. This form is widespread on sierozem and chernozem soils throughout the Riverina, the central-western and north-western plains, as well as in the central and northern slopes, and on chernozems in the upper Hunter valley, and the middle Richmond valley, on land which is flat or of less than 1 per cent. slope. (Plates I *a*, I *b*, II *a*, II *b*, III *a*).

A variation of the normal or round gilgai is found where the puffs are raised above an otherwise level surface, the shelves being flat and continuous. The diameter of the puffs ranges from 2 to 6 ft. and their surface structure is strongly granular or gunshot and 'puffy'. These have been termed 'mushroom gilgai'.

2. *Lattice gilgai.* These are intermediate between the normal or round gilgai and the wavy gilgai to be described below, and occur on slightly sloping country in the Namoi, New England, and upper Hunter regions. They can be divided into: (*a*) Those with shelf continuous and puff discontinuous and arranged in rows parallel to the direction of greatest slope; the puffs have a long axis of about 15–20 ft., of width 6 ft., and vertical interval of about 3 in. (Fig. 1 *a*); (*b*) those with the shelf discontinuous, but in rows parallel to the direction of slope, and the puffs formed not only between the shelves but across them; the shelves assume a roughly circular outline 4–8 ft. in diameter, with a vertical interval of about 3 in.

With both normal and lattice gilgai, where the puff is semi-continuous and small in area compared to the shelf, the complex may assume the appearance of a network as in Plate III *b* and Fig. 1 *b*.

3. *Wavy gilgai.* These are restricted to land which is gently sloping, the slopes varying between 15 minutes and 3 degrees. The puffs and shelves are both continuous, but are arranged in regular lines more or less at right angles to the contour, and form parallel waves which may extend

for considerable distances. The vertical interval between puff and shelf varies usually between 2 and 6 in. and has not been observed greater than 9 in. They have not been observed by the authors any farther south in New South Wales than the upper Hunter valley in the east and Dubbo in the west. They are found most frequently on the sloping

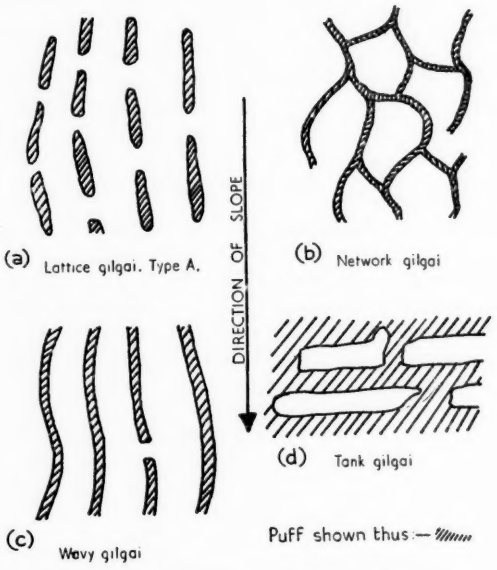


FIG. 1. Gilgai formations of sloping country.

country around the Liverpool plains, in the New England, Namoi, and Macquarie regions, and in the upper Hunter and middle Richmond valleys. In dry periods the puffs strongly resemble freshly cultivated ground, and the formation is known locally in the various districts as 'Adam's furrows', 'black-men's furrows', 'stripy country', and 'wavy country' (Fig. 1 *c*; Plates IV *a* and IV *b*).

4. *Tank gilgai.* In the central west of the State, near Forbes, the development of puffs and shelves has been observed on a large scale and in roughly rectangular form. The vertical interval is 2–4 ft., the puffs about 30 ft. wide, and the shelves 40–60 ft. long, 20–40 ft. wide, and when filled with water give the appearance of a series of tanks (Fig. 1 *d*; Plate V *a*). They appear to be quite distinct from the contour-trench formation described by McElroy (1951), Plate V *b*, from the eastern part of the tablelands.

5. *Stony gilgai.* These occur on the western plains, both east and west of the Darling river, on Stony Downs soils. The complex takes the form of those described in types 1, 2, and 3, but differs in that the whole puff is covered by stones or 'gibbers' of varying size, is extremely hard, with a

surface seal, and entirely lacks the puffy nature of the complex as originally described (Plate VI *a*).

6. *Melon-hole gilgai.* These occur under high rainfalls on the north coast of New South Wales, extending along the Richmond and Tweed rivers. The complex consists of large mounds and depressions in the centre of which are holes of varying shapes. These holes are from 3–9 ft. wide, 6–9 in. deep, and the break in the level from the depression to the hole is quite abrupt. The mounds may show abrupt subsidiary hummocks 3–6 in. high and 12–18 in. wide. They show some resemblance to the normal gilgai, but with large, abrupt sink-holes, whilst the mound is compact and massive and shows none of the crumb structure and puffy character of normal form (Fig. 2 *c*; Plate VI *b*).

Morphological Characteristics of the Profiles

The profiles described below show some of the variations as well as the constant features of the morphology of the gilgaies, but since only a very limited selection is included, they can give little appreciation of the variety encountered in the field, particularly with normal gilgaies. All sites are on land which has never been cultivated.

The descriptions are according to the Oxford system (Clarke, 1941), with the addition of some Australian terms. Thus 'self-mulching' is used to denote the condition where the dry surface soil forms quite discrete solid units of size $\frac{1}{3}$ in. and less. The structure of the lower horizons of the puff and of the subsurface horizons of the shelf often defies description in the terms of the Oxford system. To the former, consisting of pointed irregular columns of width $\frac{1}{2}$–4 in., the term 'semicolumnar' has been applied; to the latter, where the structural units are much larger, quite irregular, angular, solid and separated by fairly wide cracks, the term 'irregular polyhedral' has been applied, although the Oxford terms of 'pyramidal' and 'inverted pyramidal' are a near approach in some cases. The vertical interval between the surfaces of the puff and shelf is represented by the symbol 'v.i.' and the distance between analogous positions on adjacent hummocks or hollows is referred to as the 'wave-length' and represented by the symbol λ.

(A) *Normal or Round Gilgai*

Pallamallawa Profile (No. 160 a and b)—Black chernozemic soil

Locality: New England region, 7 miles north of Pallamallawa on edge of north-west plains.

Environment: 24 in. ann. rainfall; Pleistocene alluvium from basalt and sandstone; flat; dense tall woodland of brigalow (*Acacia harpophylla*), belah (*Casuarina lepidophloia*) and wilga (*Geijera parviflora*).

Site: Uncleared timber; puff and shelf subcircular, of roughly equal dimensions, shelf smoothly concave, v.i. 4–6 in., λ 6–10 ft.

Puff:		*Shelf:*	
0–1″	Grey chocolate clay loam; self-mulching; porous, friable; roots present but infrequent; nodules of calcium carbonate on surface.	0–3″	Black clay loam; fine small granular structure; somewhat porous, finely friable; roots present.

Puff:

1″–3″ Dull brown clay; small to medium nutty structure; porous, friable; roots infrequent; flecks of calcium carbonate.

3″–3′ Dull brown clay; medium cloddy becoming semicolumnar with depth; closed except along cracks; compact, hard; occ. roots; flecks of calcium carbonate. Merging into

3′–4′ Reddish-brown heavy clay; massive structure; closed, compact and tenacious; flecks of calcium carbonate; occ. small gravel throughout profile.

Shelf:

3″–12″ Dark brown clay; small and medium cloddy to large irregular polyhedral structure; somewhat porous, friable; roots diminishing. Merging into

1′–3′ Dark brown heavy clay; large irregular polyhedral structure; closed, compact, tenacious; flecks of carbonate. Merging into

3′–4′ Reddish-brown clay; massive structure; closed, compact, tenacious; flecks of carbonate; occ. small gravel throughout profile.

Narrandera Profile (No. 692 a and b)—Brown chernozemic soil

Locality: Murrumbidgee region, 7 m. east of Narrandera on the Wagga road.

Environment: 19 in. ann. rainfall; Pleistocene alluvium; flat; grassland of plains grass (*Stipa aristiglumis*), medics (*Medicago* spp).

Site: Open paddock; puff and shelf circular, of roughly equal dimensions, shelf smoothly concave, but with occ. sink-holes, v.i. 4–6 in., λ 10–20 ft.

Puff:

0 –2″ Grey-brown heavy clay; self-mulching; extremely porous, friable; roots frequent; nodules of calcium carbonate on surface.

2″–6″ Grey-brown heavy clay; nutty structure; porous, hard; roots present; flecks and concretions of calcium carbonate. Merging into

6″–12″ Grey-brown heavy clay; large nutty to cloddy structure; less porous, hard; grass roots infrequent; flecks of calcium carbonate.

1′–3′ Dark grey-brown heavy clay; semicolumnar to massive structure; closed except for occ. wide cracks, compact, hard; no roots; flecks of calcium carbonate; merging into

3′–5′ Grey-brown heavy clay; massive structure; closed, compact, hard; flecks of calcium carbonate.

Shelf:

0 –⅛″ Greyish-pink sandy silt; structureless or occ. laminated as a discontinuous surface seal.

⅛″–3″ Light greyish-brown heavy clay; large starchy to medium nutty structure; somewhat porous, hard, friable; roots frequent.

3″–12″ Brownish-grey heavy clay; small and medium cloddy to small and medium pyramidal and inverted pyramidal structure; occ. wide cracks; less porous except down cracks, hard, lumpy; roots infrequent; merging into

1′–5′ Brown heavy clay; large pyramidal and inverted pyramidal structure, becoming closed, hard compact below 2 ft.; no roots.

Gilgandra Profile (Nos. 51 and 52)—Brown sierozem soil

Locality: Macquarie region, 16 m. west of Gilgandra on Collie road.

Environment: 21 in. ann. rainfall; Pleistocene alluvium, of mixed origin; on top of slight ridge; savannah woodland of bimble box (*E. populifolia*), wilga (*Geijera parviflora*), *Stipa* spp., *Aristida* spp.

Site: Partly cleared paddock; puff and shelf subcircular, shelf isolated and concave, showing multiple development of smaller puffs and shelves within the unit, v.i. 2–3 ft., λ 20–25 ft.

Puff:

0 –½″ Brown loamy clay; poor crumb structure; porous, friable; roots frequent; occ. large nodules of carbonate (2″–5″ in diam.).

½″–6″ Brown loamy clay; small to medium nutty structure; porous, friable; roots frequent;

6″–12″ Grey-brown clay; large nutty to small and medium cloddy structure; occ. cracks; less porous, hard, friable; occ. grass roots present; flecks of carbonate; merging into

1′–6′ Light grey-brown clay; semi-columnar structure, becoming massive with depth; closed, compact, hard; roots not encountered; flecks of carbonate; frequent gypsum present at from 2′ 6″ to 6′; colour lightens with depth.

Shelf:

0 –⅛″ Brownish-pink sandy silt; structureless, or occ. laminated; forms a discontinuous surface seal.

⅛″–6″ Dark brown loamy clay; large crumb to large irregular polyhedral structure; occ. wide cracks; porous, lumpy; roots frequent; no visible carbonate but present by test; merging into

6″–18″ Brown clay; very large irregular polyhedral to massive structure; closed, compact; grass roots present; carbonate absent.

18″–6′ Brown heavy clay; massive structure; closed, compact; roots not encountered; carbonate flecks present but most numerous at 2 to 3 ft.; colour lightens with depth to greyish brown and yellow-brown.

(B) *Lattice Gilgai*

Murrulla Profile (No. 493 a and b)—Brown chernozemic soil

Locality: Hunter region, 5 m. south of Murrurundi on New England Highway.

Environment: 30 in. ann. rainfall; calcareous shale; undulating to rolling. Originally savannah woodland of white box (*E. albens*).

Site: Cleared paddock; puff elliptical, 5 ft. wide, 20–30 ft. long, orientated in direction of greatest slope, shelf smoothly concave, sink-holes absent, v.i. 3–4 in., λ 13 ft.

Puff:

0 –½″ Yellow-brown loamy clay; self-mulching structure; porous, friable; roots occ.; nodules of carbonate on surface.

½″–3″ Yellow-brown loamy clay; small to medium nutty structure; small vertical cracks; somewhat porous, hard, friable; roots occ.; flecks and concretions of carbonate.

3″–6″ Light brown heavy clay; changing to orange-yellow gritty clay and becoming massive with depth; vertical cracks; becoming closed with depth; hard compact; roots very occ.; flecks of carbonate.

Shelf:

0″–12″ Black heavy clay; large starchy structure; merging to small and medium cloddy structure towards 12 in.; fairly open, hard friable; grass roots abundant.

1′–2′ Black, heavy clay; massive structure; somewhat closed, compact; roots present, merging into

2′–6′ Light brown clay; massive structure; closed, compact, tenacious; roots diminishing; flecks of carbonate below 3 ft.

(C) *Wavy Gilgai*

North Star Profile—Black chernozemic soil

Locality: New England region, at foot of Blue Nobby, 13 m. east of North Star.

Environment: 25 in. ann. rainfall; Tertiary basalt; rolling, 1° 30′ slope; originally tall woodland of belah (*C. lepidophloia*), wilga (*G. parviflora*), and myall (*A. pendula*).

Site: Cleared paddock; puff and shelf both continuous, and orientated in direction of greatest slope, shelf smoothly concave, sink-holes absent, v.i. 3–9 in., λ 18–40 ft.

Puff:

0–3″ Very dark brown clay; self-mulching structure; extremely porous, hard, friable; grass roots frequent; nodules of carbonate on surface.

3″–6″ Very dark brown heavy clay; small to medium nutty structure; some wide cracks through this structure; porous, hard, friable; grass roots frequent; flecks and concretions of carbonate.

6″–18″ Very dark brown heavy clay; small and medium cloddy structure, merging into semi-columnar structure at 9 in.; vertical cracks between columns; less porous, hard, friable; grass roots present; flecks of carbonate.

18″–4′ 6″ Greyish-brown heavy clay; massive structure, but showing evidence of internal movement on the shiny surfaces exposed; closed, hard, compact; roots diminished; flecks of carbonate.

4′ 6″–6′ Reddish-brown heavy clay; massive structure; closed, hard, compact; roots absent; flecks of carbonate.

Shelf:

0–3″ Black clay; self-mulching structure; wide cracks in the lower horizons penetrate the surface and induce a rounded bank of the self-mulching horizon at their edges, extremely porous when dry, especially down wide cracks; hard, friable; grass roots abundant.

3″–2′ 6″ Black heavy clay; large starchy to small cloddy structure at the top of the horizon, merging almost immediately into very large irregular polyhedral structure; some wide cracks; closed except down cracks, hard, compact; roots very occ. Merging into

2′ 6″–6′ Black heavy clay; massive but otherwise as above; becoming reddish-brown below 4ft. 6 in.

(D) *Stony Gilgai*

Glen Lyon Profile (*Profile No. 205 a and b*)*—Stony downs soil*

Locality: Trans-Darling region, 60 m. east of Broken Hill.

Environment: 10 in. ann. rainfall; Cretaceous shales; undulating to flat; saltbush formation of saltbush (*Atriplex vesicarium*), bluebush (*Kochia spp*), predominantly in the shelf.

Site: Open gibber plains; puff and shelf subcircular, puff 18 ft. diam., shelf 6 ft. diam., shelf flatly concave with occ. sink-holes, v.i. 3–6 in., λ 24 ft.

Puff:

0″ Surface covered (about 75%) with stones ('gibbers') of size ¼–3 in., predominantly 1–2 in., mainly quartzite, some ferruginous (see Plate V *a*).

Shelf:

0–½″ Orange-brown sand; occ. gibbers; structureless or laminated, forming a discontinuous surface seal; wide cracks present; closed except for cracks; compact.

Puff:

0 –1″	Brown sandy loam; laminated, forming a continuous surface seal except between gibbers; closed, compact, very occ. cemented at the junction with the underlying horizon.
1″–6″	Reddish-brown clay; small cubic structure, the cubes from 1–3 in. very well defined, of size ≏ ⅛ in., and arranged in pointed columns; below this size increases with depth; open only down inter-columnar spaces; hard; roots present.
6″–12″	Reddish-brown sandy clay; medium and large nutty to semicolumnar structure; open only down intercolumnar spaces; compact but easily broken; some roots present, merging into
1′–6′	Reddish-brown light clay; semicolumnar structure becoming massive with depth; closed, compact; roots few; flecks of carbonate below 1 in.; colour lightens to orange-brown with depth; gibbers absent below 1 in.

Shelf:

½″–3″	Light reddish-brown silty sand; crumb microstructure, large irregular polyhedral macrostructure, with horizontal laminations (representing successive wind deposits?); wide cracks; porous, almost free running; frequent clumps of roots; no visible carbonate, but present by test.
3″–2′	Light reddish-brown sandy clay; very large irregular polyhedral structure, coarsening to massive with depth, and with horizontal laminations across units; wide cracks at the top of the horizon; closed, hard, compact but powdery when broken; roots present; no visible carbonate, but present by test.
2′–6′	Orange-brown sandy clay; massive structure; closed, compact; roots infrequent; flecks of carbonate; occ. gibbers throughout profile.

(E) *Tank Gilgai*

*Forbes Profile (No. 840 a and b)—Brown sierozem soil**

Locality: Lachlan region, 5 m. north of Fobes on Gunningbland road.

Environment: 19 in. ann. rainfall; shale, locally undulating; originally shrub woodland of belah (*C. lepidophloia*) and bimble box (*E. populifolia*).

Site: Cleared paddock; shelf sub-rectangular, partially submerged, 10–12 ft. wide, 60–90 ft. long; puff 30 ft. wide, v.i. 1 ft. 6 in. to 3 ft.

Puff:

0 –⅛″	Light brown silty sand; structureless, occ. laminated forming a discontinuous surface seal; occ. nodules of carbonate; occ. stones.
⅛″–1″	Brown loamy clay; crumb structure; very porous, friable; roots frequent; occ. nodules of carbonate; occ. stones.
1″–6″	Brown loamy clay; large crumb to medium nutty structure, coarsening with depth; somewhat porous, hard, friable;

Shelf:

0 –⅛″	Light brown silty sand; structureless, occ. laminated; forms a discontinuous surface seal.
⅛″–3″	Grey-black loamy clay; large crumb to small cloddy structure; porous, friable; roots frequent.
3″–9″	Greyish-black clay; medium to large cloddy structure; somewhat porous, slightly compact; roots present.
9″–6′	Greyish-brown heavy clay; massive structure; closed, com-

* Profile description by S. A. Waring

Puff:

roots present; occ. flecks and concretions of carbonate.

6″–1′ 10″ Light brown heavy clay; semi-columnar to massive structure; somewhat closed, slightly compacted; grass roots infrequent; numerous flecks of carbonate.

1′ 10″–6′ Greyish-brown clay; massive structure; closed, compact; roots not encountered; carbonate absent, gypsum crystals present in large quantity between 2 ft. and 3 ft. 5 in., but not below this; colour alters with depth to grey mottled with orange and black at 5 ft.

Shelf:

pact; roots infrequent; flecks of carbonate from 9 in. to 16 in., but gypsum absent; colour alters with depth to grey with slight orange mottling.

(F) *Melon-Hole Gilgai*

Gundarimba Profile (No. 494 a and b)—Meadow soil

Locality: Richmond-Tweed region, 6 m. south of Lismore on Coraki road.

Environment: 53 in. ann. rainfall; recent river alluvium from basalt; flat valley floor in hilly country; originally meadow.

Site: Open paddock; puff and shelf sub-circular to irregular, shelf with sudden, abrupt, sink-hole, 6 in. deep, 5 ft. wide, and puff with sudden, abrupt mound, 3 in. high, 15 in. wide, these mounds occurring with greatest frequency where land has been cultivated and gilgaies reforming. λ 20 ft.

Puff:

0–6″ Dark grey heavy clay; massive or pyramidal structure; occ. cracks; closed, hard, compact; with mat of grass roots.

6″–6′ Dark grey heavy clay; massive structure; closed, hard, compact, tenacious, wet; roots diminish throughout; red-brown flecks of Fe_2O_3 in lower horizons.

Shelf:

0–¼″ Dark brown silty clay; fine crumb structure; scattered deposits of iron oxide on surface, apparently associated with holes left by old rush stems.

¼″–6′ Very dark grey heavy clay; massive structure; grass roots abundant diminishing in lower levels; closed, compact, very tenacious; colour lightens with depth to grey; red-brown flecks of Fe_2O_3 throughout.

The Morphology of the Gilgai Formations

The features of the profiles show that whilst the shelves generally have the morphological features of their soil group, the surface soil of the puff shows some resemblance to the subsoil of shelf profile. This is particularly the case in those found on chernozemic or sierozem soils, where carbonate is almost invariably present at, and gypsum comes closer to, the surface in the puff than in the shelf. Indeed, in some cases, large gypsum crystals have been found on the surface of the puff also (e.g. at Gilgandra and Mungindi). Accordingly it might be postulated that the puff represents part of the subsoil which has been forced to the surface, as suggested earlier (cf. Prescott, 1931; Leeper, 1949). Further evidence of this is obtained from the selected examples described below, which give some idea of the variations in the morphology of the gilgai formation.

In the Riverina the most widespread type is the normal or round gilgai, in which the yellow-brown friable and crumb-structured clay of the puff is in marked contrast to the tenacious compact but slightly cracked grey clay of the shelves. Carbonate is invariably present on the puff and absent from the shelves. A pit cut through to about 4 ft. deep from one puff to the next shows that the depth at which this yellow-brown clay appears

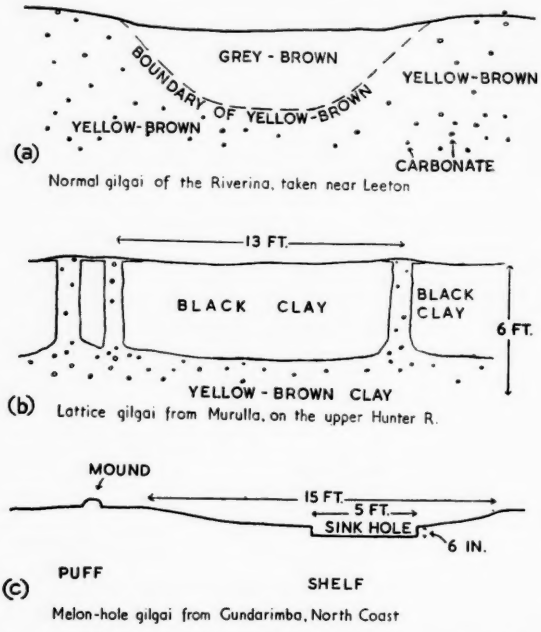


FIG. 2. Diagrammatic representation of the cross-section of certain gilgai formations.

increases regularly away from the puff, and is not encountered until about 3 ft. below the surface of the soil at the mid position between the puffs (Fig. 2 *a*). This type has already been described by H. N. England (Prescott, 1931), A. Howard (1939), and others.

A particular variant of this found over a limited area on the Yenda-Rankins Springs stock route has been formed on colluvium from basalt and contains numerous stones in the profile. A cross-section* shows that not only does the carbonate go down below the surface of the shelf but that the stone line does the same (Plate II *b*). Another feature shown here was the crack pattern, cracks on the side of the puff, which were quite invisible on the surface, widened with depth, and one crack in particular widened sharply below 2 ft. to be 10 in. wide at 5 ft. Although

* Cut by courtesy of H. N. England and the Water Conservation and Irrigation Commission of New South Wales (Leeton).

a similar occurrence of stones in the gilgai has been found near Narrabri, such formations are rare in eastern New South Wales and serve to relate these gilgaies to the stony gilgai formations found west of the Darling river.

In one complex of gilgaies examined near Gilgandra (Profiles 50 and 51) the whole unit had a diameter between 20 and 25 ft., the puffs varying from 2 ft. to 2 ft. 9 in. above the level of the shelves. These showed multiple development, with up to 3 or 4 separate holes within one basin or shelf. At the time of sampling cracks up to 2 in. across the top occurred along the arcs of circles at varying distances from the centre of the depression. In each of the holes at this site the piece of soil of the surface could be seen to have broken away and have fallen some way down the hole. From the profile descriptions it can be seen that the gypsum horizon is much closer to the surface in the puff than in the shelf, whilst massive concretions of pebbles cemented by calcium carbonate (up to 6 in. diameter) were found on the centre of the puffs.

Another area of gilgaies near Gilgandra that had been graded level and cultivated for wheat was examined in October 1950, after very high rainfalls in 1949 and 1950. During these 2 years heaving of the surface soil had been particularly marked, and it was noticeable that, scattered across the surface, relatively thin layers of light brown clay overlaid areas of black topsoil. This light brown clay was similar in character to that encountered 3–5 ft. below the surface of the shelf or 1 ft. 6 in. below the puff. It was distributed as small 'flows' mostly 2 to 3 ft. long, each spread in a tongue from 6 in. up to just over 1 ft. wide. When viewed they were never more than 3 in. thick, and usually much thinner.

The Murulla gilgaies (Profiles 493 *a* and *b*) were in savannah woodland that had never been cultivated. A pit dug across between these two profiles showed the yellow-brown soil similar to that found in subsoil of the shelf, coming up to the surface of the puff in a definite column 9 in. wide (Fig. 2 *b*). The contrast in colours between the almost black clay of the shelf and the bright yellow-brown of the puff was very striking, since the division between them was sharp. Immediately adjacent to the puffs, the almost black clay of the shelf was covered by a thin layer of yellow clay, which was sharply demarcated from it. This appeared to have washed off the puff.

At one site at North Star the wavy gilgaies were examined in greater detail. Here the land had a practically constant slope of 1 degree 30 minutes in an east-north-easterly direction. The wave-length was 21 ft. and the vertical interval 8 in. At the upper end where the slope became steeper both wave-length and vertical interval showed some variation, and puff and shelf lost their identity when the slope became steeper than about 2 degrees 30 minutes. A pit cut at right angles to the complex (i.e. along the contour) from one puff, through the shelf and to the other puff showed the following changes.

In the shelf the topsoil was a black self-mulching layer 1½–2½ in. thick, rich in organic debris and roots. It terminated suddenly in a small cloddy layer which formed the upper part of poorly defined, jointed, and angular columns of small but varying cross-section, extending to a depth of

2 ft. 6 in. The vertical or near vertical cracks between the columns were never wider than about 1 in. Below about 4 ft. 6 in. the veins of the black soil formed a network of fine intrusions extending into the dull red clay of the lower horizons, even into that part of the soil below 5 ft. which was not obviously cracked at the time of sampling, indicating a greater development of cracking during drier periods.

Except for a dark brown zone at the centre of the puff, the surface of soil was similar in colour to the upper layers of the shelf. Lower down the profile the brown colour spread out, grading into grey-brown between 2 and 3 ft., through reddish-brown at 4 ft. 6 in. into the dull red of the lower horizons of the shelf below 5 ft. The surface had a very well-developed small crumb structure, was highly friable, and contained small concretions of calcium carbonate. Below this the structural units increased in size through medium nutty to medium and large cloddy masses arranged in rough columns. Where the grey-brown colour spread out, below about 2 ft., the structural units also changed, and showed a series of curved cracks, spreading farther outwards with increasing depth and becoming almost horizontal about 3 ft. from the centre of the puff. The shape of these units is rather similar to the 'amygdaloidal' units described by Krishna and Perumal (1948) from the black cotton soils of the Deccan.

The melon-hole gilgais found under higher rainfalls on the lower Richmond river show certain differences from the other forms. They are usually acid throughout the profile and contain no carbonate. The shelf contains from one to three smaller holes, roughly circular, ranging in size from 3 to 6 ft. in diameter. Sometimes the holes are elongated into a bean shape. For much of the year the holes are filled with water and reeds and rushes grow in them. Much of this country has been graded level for pasture establishment, as it is usually too heavy for cultivation, and the melon-holes re-establish themselves over a period of years, or more rapidly if flooding occurs. After the severest drought on record (1951–2) the subsoil remained moist below about 3 ft., even though wide cracks occurred in the hole, across the shelves, and adjacent to the puff. (J. McGarity, priv. comm.).

Chemical and Physical Characteristics

Analyses have been made on sixty profiles of gilgai soils in an attempt to find what properties they possess in common and which at the same time differentiate them from adjacent soils not showing this phenomenon. Most of the profiles examined, and all of the comparisons with non-gilgaied soils, come from the chernozemic and sierozem soils, since in these the gilgaies showed the greatest variety.

Mechanical composition

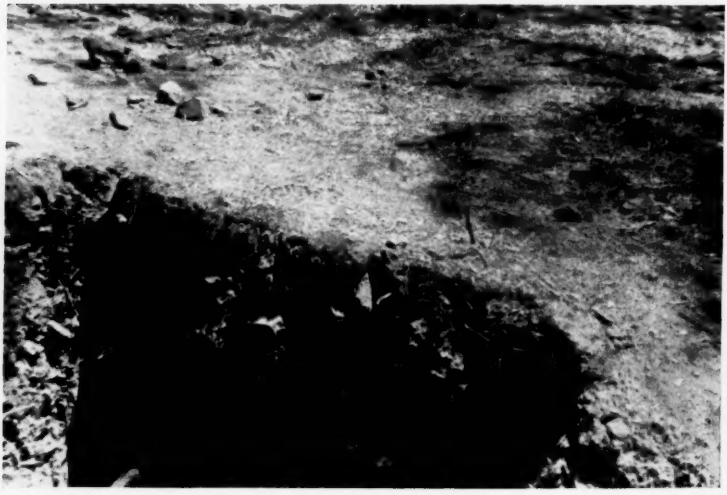
In both chernozemic and sierozem soils the normal profile is a medium to heavy clay (40–60 per cent. clay), with a gradual increase in clay content with depth. The gilgaied variants of both groups are distinctly coarser in texture, and the shelf profile in each case shows a marked increase in clay content with depth, ranging from 30–40 per

a. Normal gilgai near West Wyalong. The lighter areas are the shelf with a more profuse vegetative cover. *Photo.: S. A. Waring*

b. Normal gilgai from the Pilliga, carrying brigalow, belah, and box (*Acacia harpophylla, Casuarina lepidophloia, Eucalyptus pillagaensis, E. populifolia*). The vertical interval between puff and shelf is 3 ft. *Photo.: H. D. Waring*

E. G. HALLSWORTH, G. K. ROBERTSON, F. R. GIBBONS—PLATE I

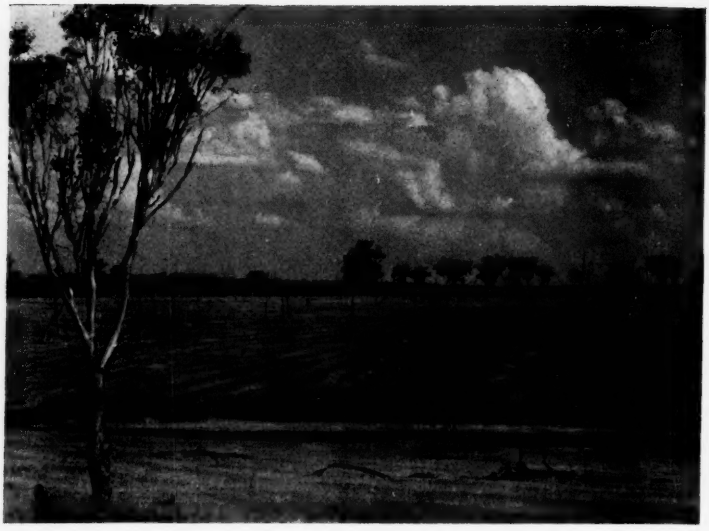
a. Acid gilgai from the Pilliga, carrying brigalow (*A. harpophylla*). The sharp rise of the puff is particularly noticeable: the vertical interval here is 4 ft. *Photo.: S. A. Waring*



b. Section through a normal gilgai carrying some stones, north of Yenda. The stones can be seen on the surface of the puff on the right, and going down under the shelf to the left foreground. White flecks of carbonate can be seen below the stone line. Note the crack in the left foreground, which expanded to 10 in. wide at 5 ft. deep. *Photo.: E. G. Hallsworth*

a. Normal gilgai from Goolgowi, with mallee (*E. dumosa*, *E. oleosa*) growing on the puffs only. A crack that has become a hole can be seen in the foreground. *Photo.: P. H. Walker*

b. Network gilgai, North Star. The slightly raised puffs form the network. *Photo.: J. Bunt*



a. Wavy gilgai on chernozemic soils on the Delungra–Yetman road. The shelves appear lighter in colour. The varied length of the puffs can be seen on the right centre. *Photo.: J. D. Colwell*

b. Wavy gilgai, Musswellbrook, Hunter valley. Photographed in a dry season. The darker lines are the puffs, almost bare of vegetation. *Photo.: B. W. C. Fox*

a. Tank gilgai, 10 miles south of Bogan Gate, photographed after rain. *Photo.: S. A. Waring*

b. Contour trench formation, Happy Jack's Plain, Monaro. These form on silty bog soils in isolated pockets close to the tree line. They are coarse textured and high in organic matter. The trenches are generally narrow and run along the contour. (*McElroy 1952*)

a. Stony gilgai, from Mt. Stuart. In most years the stone covered puffs are bare, as seen here, and the vegetation is confined to the shelves. *Photo.: W. E. Darley*

b. Melon-hole gilgai near Lismore. These are filled with water for most of the year, except where the land has been drained. *Photo.: J. D. Colwell*

Surface cracks in a slightly gilgaied soil near Inverell. *Photo.: J. D. Colwell*

cent. clay in the surface to 55–65 per cent. clay in the subsoil. The surface soil of the stony gilgaies, with only 10–30 per cent. clay, is much coarser in texture, whilst the melon-hole gilgaies are the heaviest textured soils encountered, with 60 per cent. and 90 per cent. of clay in the surface and subsoil respectively. Analyses of individual profiles show that the surface soil of the puff is normally but not always richer in clay than that of the shelf.

Some association is shown between the mechanical composition and

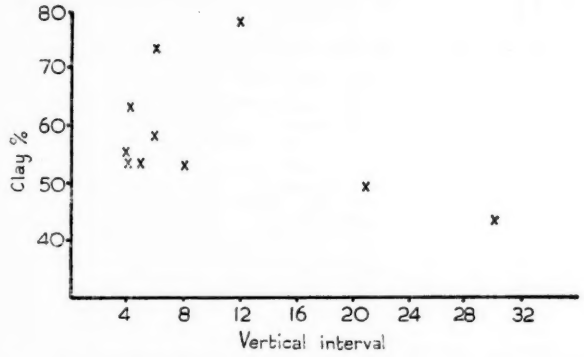


FIG. 3. Mean clay content and vertical interval for some gilgai soils.

the vertical interval of the gilgai (Fig. 3). The higher the clay content, the smaller is the vertical interval between puff and shelf. There appears to be no similar relationship with the wave-length.

Clay composition

All the gilgai soils contain montmorillonite and kaolinite, commonly with some quartz, whilst the micaceous clay minerals are frequently present in the sierozemic soils formed on alluvium, particularly in the Riverina. The clay mineral pattern varies considerably, and no constant differences are found between the normal soils and their gilgaied variants. Selected results are given in Table 1.

In the gilgaied chernozemic and sierozem soils kaolinite ranges from 20 to 40 per cent. and montmorillonite from 30 to 80 per cent. In all the shelf profiles the proportion of montmorillonite increases with depth, the increase varying from 3 to 30 per cent., but the puff profiles show no regular change.

Calcium carbonate

A feature of the gilgai soils that has attracted attention in the past is the presence on the surface of nodules of calcium carbonate. These are commonly found on the puff but are absent or very infrequent on the shelf. More extended field studies show that carbonate is not invariably present. It is completely lacking, for example, in the normal gilgaies of

TABLE I

X-Ray and Thermal Analysis of Clays from some Gilgai Soils

Soil No.	Locality		Depth	Exch. cap./100 g. clay	Composition: X-ray analysis: Montmorillonite	Kaolinite	Illite	Quartz	Thermal analysis: % montmorillonite	% kaolinite	Chemical analysis: % illite*
	Stony Gilgaies:										
226.A.10	Corona, Broken Hill.	Shelf	5–6′	18	+++	++	..	+	n.d.		n.d.
222.A.10	Mokeley. Tibooburra.	,,	4–5′	40	+++	++	..	..	n.d.		n.d.
205.A.11	Box Tank, Menindee.	,,	5–6′	28	+++	+	..	..	n.d.		n.d.
	Gilgaied Sierozem:										
49.1	2 miles west of Gilgandra.	Puff	0–2″	94	++++	++	++	++	55	37	6
49.7	,, ,,	,,	5–6′	86	++	++	++	+++	57	30	10
50.1	,, ,,	Shelf	0–2″	60	+++	+	++	+++	42	25	8
50.7	,, ,,	,,	5–6′	93	++++	+	+	+++	53	35	6
51.1	Berida Station, Gilgandra.	Puff	0–2″	55	+	++	++	++	30	20	24
51.5	,, ,,	,,	3–4′	62	+	+	+	++	40	40	16
52.1	,, ,,	Shelf	0–2″	72	+++	++	++++	+++	54	22	25
52.5	,, ,,	,,	3–4′	78	+++	++	+	++	57	24	18
496A.1	Narrandera.	Puff	0–3″	n.d.	+	++	+++	+	n.d.		n.d.
	Gilgaied Chernozemic:										
115.1	Kelvin. Liverpool Plains.	Puff	0–3″	53	+	++	++	++	25	47	28†
115.6	,, ,,	,,	2–3′	41	n.d.				27	47	21†
116.1	,, ,,	Shelf	0–3″	36	+	++	++	+	23	23	54†
116.7	,, ,,	,,	2–3′	48	++	++	+++	++	53	28	20†
160B.1	Pallamallawa. NW. slopes.	Puff	0–3″	73	++	+	..	++	56	43	n.d.
160B.6	,, ,,	,,	2–3′	60	++	+++	..	++	68	29	n.d.
160A.1	,, ,,	Shelf	0–3″	70	++	++	..	++	51	45	n.d.
160A.6	,, ,,	,,	2–3′	58	+++	+	..	++	71	29	n.d.
43.1	Scone. Upper Hunter Valley.	,,	0–3″	88	+++	++	+	++	48	25	20
43.5	,, ,,	,,	3–5′	123	++++	+	+	++	57	25	9
	Gilgaied Meadow Soils:										
494A.5	Gundarimba. North coast.	Puff	2–3′	n.d.	n.d.				60	40	n.d.
494B.1	,, ,,	Shelf	0–3″	34	+	++++	..	..	32	72	n.d.
494B.5	,, ,,	,,	2–3′	30	+++	++	..	..	64	31	n.d.

The X-ray photographs were taken using glycerol-treated aggregates after MacEwan (1949). The method of thermal analysis will be described in a later paper.

* Calculated from the K_2O content of the clay, assuming that illite contains 5% K_2O as a maximum (Alexander, Hendricks, and Nelson, 1939).

† Determined by difference.

the once lateritized Tertiary alluvials of the Hawkesbury valley, and in the acid gilgaies found in parts of the Pilliga. On the north coast the melon-hole gilgaies of the meadow soils of the high rainfall areas are quite devoid of carbonate, although it is present in the deep subsoil of the gilgaied chernozems of 40–50-in. rainfall areas. It is almost invariably present at the surface of the puff of the gilgaied chernozemic and sierozem soils, whilst in the drier areas of the Stony Downs carbonate may be found in the top half-inch or so of the shelf (Table 2).

TABLE 2

Soil Reaction and Calcium Carbonate Content of Some Chernozemic and Sierozem Soils

	$CaCO_3$ (%)			*Soil reaction* (pH)		
	Gilgaied		*Non-*	*Gilgaied*		*Non-*
Depth	*Puff*	*Shelf*	*gilgaied*	*Puff*	*Shelf*	*gilgaied*
			Chernozemic			
0–3″	3·0	0·2	0·3	7·6	6·5	6·9
3–6″	4·5	..	0·5	7·8	6·5	7·0
6–12″	5·3	..	0·7	7·9	6·5	7·0
1–2′	6·4	0·3	2·0	8·0	7·3	7·2
2–3′	6·8	2·3	3·8	8·1	7·8	7·7
3–4′	7·1	2·5	6·1	8·0	7·8	7·9
4–5′	2·0	1·6	7·4	7·6	7·9	8·1
5–6′	0·3	0·2	7·0	7·5	7·6	8·1
			Sierozem			
0–3″	2·7	0·2	1·2	7·8	6·3	8·1
3–6″	4·5	0·8	1·5	8·0	6·7	8·3
6–12″	8·4	3·1	1·9	8·1	7·1	8·2
1–2′	8·8	3·4	1·5	8·2	7·7	..
2–3′	8·2	2·5	1·3	8·3	8·0	8·1
3–4′	4·0	1·1	0·8	7·9	8·0	8·2
4–5′	3·7	1·0	0·7	7·9	7·8	..
5–6′	3·0	1·0	0·9	7·9	7·5	7·9

Note: No. of profiles analysed: Gilgaied chernozemic, 11; Gilgaied sierozem, 4
Normal chernozemic, 6; Normal sierozem, 9

Soil reaction

A difference in reaction is found between the surface soil of puff and shelf which follows the differences in carbonate distribution, and the range of pH values for the surface 3 in. is shown in Fig. 4. The differences between the values of puff and shelf are greatest for the chernozemic soils, less for the sierozem, and least for the stony gilgaies. The relationship of these differences to the rainfall is shown in Fig. 5.

Exchangeable cations and exchange capacity

The determination of exchangeable cations from forty profiles (after Black, 1947) showed that calcium is the principal cation in the surface horizons of puff and shelf of all soil groups, and accounts for more than 70 per cent. in the chernozemic and sierozem soils. The proportion of

* Calculated from the K_2O content of the clay, assuming that illite contains 5% K_2O as a maximum (Alexander, Hendricks, and Nelson, 1939).
† Determined by difference.

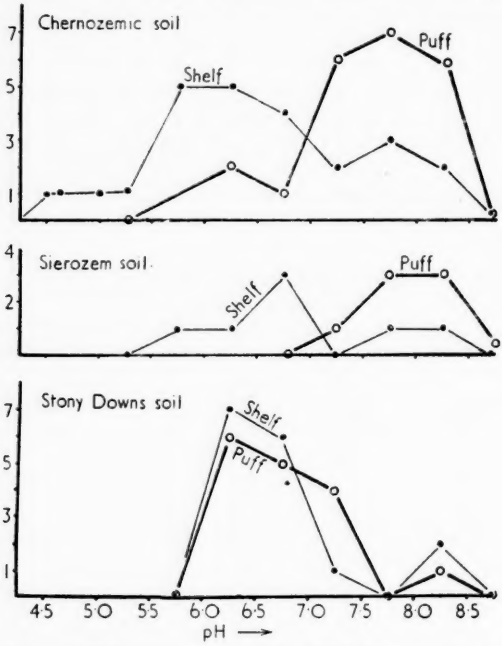


FIG. 4. pH-frequency curves for gilgaied chernozemic, sierozem, and stony downs soils.

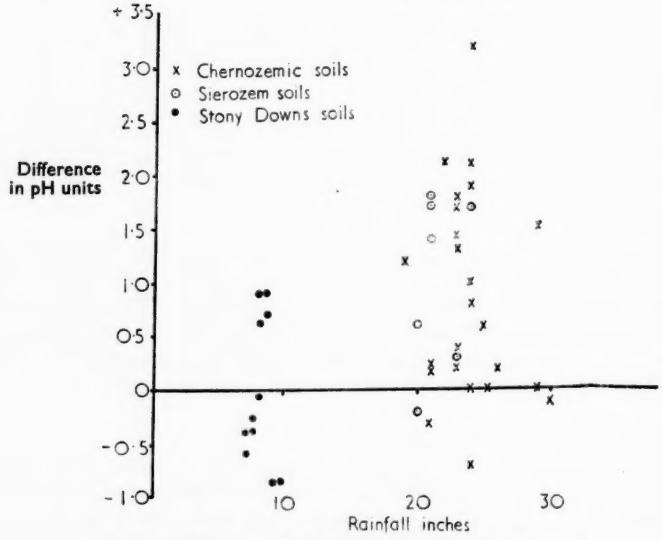


FIG. 5. Difference in pH between puff and shelf in relation to rainfall.

513.6.1

TABLE 3

Exchangeable Cations of Some Gilgaied Soils (in m.e. per 100 g. of soil)

(*a*) *Chernozemic Soils*

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Normal gilgai. Nr. Mussellbrook, Hunter Valley. Shelf (Profile 39)					Puff (Profile 40)				
0–4″	22·8	4·0	2·0	1·0	35	43·4	10·3	1·4	0·9	56
4–12″	18·9	6·4	1·4	1·6	38	38·1	14·6	0·2	1·7	55
18–24″	38·7	12·0	1·1	6·7	58	23·7	18·7	0·3	5·3	48
2′ 6″–3′ 0″	20·9	13·6	0·9	6·6	42	18·1	19·4	0·4	7·1	45
4′ 0″–4′ 6″	37·1	17·0	0·5	10·4	65	21·6	20·0	1·7	8·7	52

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Wavy gilgai. Kelvin, Liverpool Plains. Shelf (Profile 116)					Puff (Profile 115)				
0–3″	11·3	2·9	0·7	0·6	21	23·6	5·0	0·2	1·2	30
9·12″	14·0	7·4	..	0·8	31	17·0	8·6	..	1·4	27
1–2′	15·9	11·1	..	1·1	36	11·5	10·3	..	2·2	24
2–3′	16·7	11·4	..	1·9	30	13·9	8·7	..	2·4	25
4–5′	14·8	11·1	..	2·1	28	14·8	7·0	..	3·2	25
5–6′	13·5	10·1	..	1·4	25	18·5	6·9	..	2·6	28

(*b*) *Sierozem Soils*

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Normal gilgai. Berida Station, 17 m. W. of Gilgandra. Shelf (Profile 52)					Puff (Profile 51)				
0–2″	23·2	2·2	2·5	0·4	37	19·0	1·8	1·0	0·2	22
2–12″	28·5	2·5	1·2	0·8	33	19·3	4·8	1·0	0·9	26
1–2′	18·4	2·4	0·9	2·3	24	14·0	8·3	1·0	2·7	26
2–3′	9·5	2·7	1·2	6·6	20	6·7	7·3	0·9	4·1	19
3–4′	14·7	3·8	0·8	6·7	26	17·8	6·4	0·9	5·9	31
6–7′	13·2	3·9	0·8	9·1	27	←——n.d.——→				

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Normal gilgai. 2 m. W. of Gilgandra. Shelf (Profile 50)					Puff (Profile 49)				
0–2″	17·7	2·6	0·9	0·4	28	33·5	2·6	0·6	0·3	37
2–12″	23·2	3·2	0·7	0·6	34	33·3	5·3	0·9	0·5	40
1–2′	28·0	4·8	0·8	1·1	36	33·3	7·7	0·4	0·6	42
2–3′	40·5	5·1	1·6	1·8	49	34·1	8·9	1·3	0·7	45
3–4′	35·1	5·8	1·3	1·8	44	30·9	8·4	1·3	1·3	41
4–5′	33·0	6·7	0·7	1·6	42	27·4	8·2	1·1	1·3	38
5–6′	30·8	4·6	1·7	1·9	39	30·0	8·2	1·4	2·4	42

(*c*) *Stony Gilgaies*

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Box Tank, Glen Lyon. Broken Hill. Shelf (Profile 205 B)					Puff (Profile 205 A)				
0–½″	2·3	0·6	0·1	0·1	8	←——n.d.——→				
0–1″	←——n.d.——→					18·0	3·4	1·4	0·2	23
1–3″	11·2	1·3	0·6	1·5	19	21·0	6·6	0·7	2·7	31
5–6′	16·3	2·8	0·8	9·1	29	13·0	5·2	0·6	9·2	28

Depth	Exchangeable Ca	Mg	K	Na	Exch. cap.	Exchangeable Ca	Mg	K	Na	Exch. cap.
	Nr. Mokeley Tank. Tibooburra. Shelf (Profile 222 A)					Puff (Profile 222)				
0–½″	11·9	3·1	1·2	0·6	18	←——n.d.——→				
0–1″	←——n.d.——→					6·7	5·9	0·6	0·8	13
1–3″	12·3	3·1	1·7	1·9	19	3·6	13·3	0·9	7·2	25
5–6′	18·3	9·5	0·3	11·9	40	16·4	10·9	0·3	13·4	41

exchangeable magnesium and sodium always rises in the lower horizons. Exchangeable potassium is never high and the amount decreases with depth. Some typical results are given in Table 3.

The mean results for gilgaied and normal chernozemic soils (Table 4) show that whilst both variants show an increase in exchangeable sodium and magnesium with depth, the trend is much more marked with the gilgai soils. In particular the higher proportion of sodium is probably significant, since de Sigmond (1937) showed that the dispersive effect of the sodium ions becomes appreciable where they exceed 10–15 per cent. of the exchangeable cations.

TABLE 4

The Exchangeable Bases of Normal and Gilgaied Chernozemic Soils

Depth	*Normal* (*mean of 6 profiles*)				*Gilgaied* (*mean of 11 profiles*)			
	Ca	Mg	K	Na	Ca	Mg	K	Na
	(*as % of total exchangeable bases*)							
0–4″ . .	59	34	2	4	83	11	4	3
4–12″ . .	58	36	1	4	77	16	3	3
1–2′ . .	50	42	1	6	64	25	2	9
2–3′ . .	50	42	1	6	55	28	2	15
3–4′ . .	44	44	1	8	54	26	2	17

In all gilgai soils examined the cation exchange capacity of the clay fraction increases with depth (in the shelf). This accords with the distribution of montmorillonite, and in fact appears to be largely a function of the montmorillonite content, for if M, K, and I are the percentages of montmorillonite, kaolinite, and illite respectively, the regression equation obtained is

$$\text{Exchange capacity}^* \text{ of the clay} = 0{\cdot}99\,M + 0{\cdot}02\,K + 0{\cdot}18\,I,$$

only the coefficient for montmorillonite being significant ($p < 0{\cdot}001$).

Soluble salts

The soluble salt contents were measured for several profiles but appear unrelated to the phenomenon. The figures range from 0·06 to 2 per cent., slightly falling in the immediate subsurface horizons and rising to higher values in the subsoil. No differences are shown between the gilgaied and normal soils, nor are there any consistent differences between puff and shelf profiles.

The Factors affecting Soil Swelling

From the analyses described above it follows that the acidity, carbonate content, and soluble salts content cannot be considered to show any causal relationship to the gilgai phenomenon. It can be seen, however, that the clay content, the proportion of montmorillonite in the clay, and the proportion of sodium on the exchange complex increase with depth in

* Calculated free from organic matter, using the equation Exchange capacity = 0·82 × % Clay + 5·12 × % Organic Carbon, the derivation of which will be published elsewhere.

the shelf, although not all profiles show increases in all three. Since changes in each of these three factors might be expected to affect the expansion of the soil on wetting, the volume expansion of the different horizons of various gilgaied and non-gilgaied soils was measured using the modified Keen-Raczkowski method* (Coutts, 1930).

The volume expansion patterns for the shelf profiles of gilgaied chernozemic and sierozem soils are similar and the average swelling capacity increases from surface to subsoil by about 15 per cent. The expansions of the subsoil of the puff and the shelf are similar, but those of the surface soils are variable. In those soils in which the clay content of the puff is lower than that of the shelf the swelling is lower, but where the clay content is higher the swelling of the puff is greater than that of the shelf. Since on the field evidence the puffs represent subsoil which has been brought to the surface, for comparison with other soil groups the swelling pattern of the shelves only will be considered. They are illustrated in Fig. 6.

From these results it can be seen that the swelling patterns for gilgai soils, whether they occur as variants of meadow, chernozemic, sierozem, or stony desert soils, show exactly the same trend to higher values in the subsoil. The increase in the swelling capacity of the subsoil over that of the topsoil varies from 15 to 30 per cent. and is significant in all cases. A comparison of swelling pattern of gilgaied and non-gilgaied forms shows that the normal chernozemic and sierozem soils exhibit a high swelling capacity (30–40 per cent.) with but little change down the profile, which is in contrast to the marked increase from surface to depth found in the gilgaied forms.

In all the non-gilgaied soils in which there is a similar increased swelling with depth, viz. solodic soil (Hallsworth, Costin, and Gibbons 1953), solodized-solonetz, red-brown earth, chocolate soil, and euchrozem (Hallsworth, Colwell, and Gibbons, 1953), there is no marked surface cracking. Except for the euchrozems, the swelling is of an altogether lower order of magnitude (less than 10 per cent. at the surface). The failure of the euchrozems to develop gilgaies is probably due to the high content of free sesquioxides, for a fine crumb structure develops on drying, without the formation of large, wide cracks.

The only other group examined which shows surface cracking, namely the clay pan soils of the West Darling region, shows no significant increase in swelling with depth, the topsoil and subsoil swelling under the same moisture conditions to much the same extent.

The differential swelling pattern of gilgai soils is consequently quite characteristic. The subsoil swells more than the topsoil, the increase ranging from 15 to 30 per cent., whilst the swelling of the topsoil is always more than 10 per cent., so that under suitable conditions of moisture change the swelling and shrinkage are sufficient to produce surface cracking.

The swelling capacity of a soil on wetting is due to the quantity and

* Since the expansion of some of the heavier soils was quite considerable it was found necessary to wrap a strip of filter-paper held by a rubber band around the outside of the box to prevent loss from spilling when the soil was wetted.

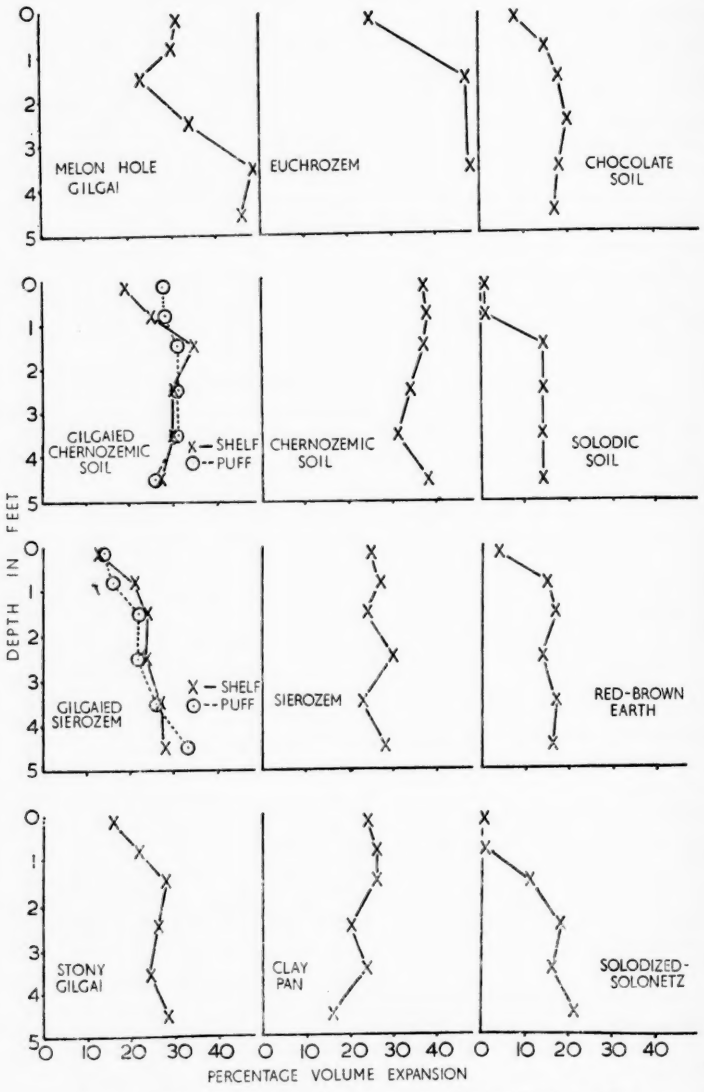


FIG. 6. The swelling profile of certain gilgai and other soils.

character of the colloids present, both organic and inorganic, and to their hydration capacity. For the inorganic colloids this varies with the clay mineral present and with the nature of the cations on the exchange complex. Restricting consideration to the chernozemic and sierozemic soils and their gilgaied variants, the relation found was:

$$V = 0{\cdot}47\,C + 0{\cdot}70\,\mathrm{Na} + 0{\cdot}24\,E + 0{\cdot}62\,H,$$

where V = percentage volume expansion,

C = percentage clay,

Na = m.e. exchangeable Na/100 g. soil,

E = Exchange capacity of the clay in m.e. per cent.,

H = percentage organic carbon (Walkley-Black).

All except organic carbon were significantly related to the volume expansion. Recalculating the data, but omitting the organic carbon, gave the equation

$$V = 0{\cdot}48\,C + 0{\cdot}76\,\mathrm{Na} + 0{\cdot}24\,E.$$

The coefficient for C was significant at the 0·1 per cent. level, that for Na at the 1·0 per cent., and that for E at the 0·1 per cent. level. (For the whole equation $r = 0{\cdot}79$ and $p < 0{\cdot}001$.)

Since for these soils the exchange capacity of the clay has already been shown to be significantly related to the amount of montmorillonite present, it can be postulated that in the gilgai soils (shelves) examined the clay content, montmorillonite content, and exchangeable sodium, all of which increase with depth, may be held responsible for the increased swelling capacity of the subsoil over that of the topsoil.

Discussion

The undulations of the soil surface described by workers outside Australia do not appear to be completely analogous to the gilgaies. Thus the hummocks of the Red River valley in Canada are considered by Ellis and Shafer (1928) to be due to material from the more greatly swelling surface soil falling down cracks in the sandy clay beneath. The 'hog-wallow' soils of the United States have been attributed to various agencies, but no conclusive explanation has been offered (Nikiforoff, 1941), whilst the 'quagmires' of India have been shown (Krishna and Perumal, 1950) to be due to the pressure of fluid soil. Field observations in New South Wales show that the 'quagmire' explanation cannot hold here, whilst the volume expansion figures given show the gilgaies to be quite different from the Red River valley hummocks, which have sandy subsoils. Within Australia, gilgaies have been described several times and probable mechanisms for their formation discussed, although no serious analytical studies have been made. Some of the suggestions advanced, however, have been largely substantiated by the present work. In the first general account of the phenomena, Prescott (1931) suggested that the alternations of wet seasons (or floodings) with droughts would induce the characteristic undulations. These undulations, according to work in the Mt. Gellibrand area of Victoria, occurred only in patches in swampy

areas (Leeper *et al.*, 1936). For their formation Leeper considered that the soil should contain less than 50 per cent. sand, and that the swamps should be areas of internal drainage. The mechanism he suggested was that when the soil dries, pieces of surface material fall down the cracks. Then when the soil is re-wetted and swells, an increased pressure will be developed at these points exerted sideways and upwards, leading to the fracture of blocks of soil that are ultimately forced to the surface. From observations in the Riverina, Howard (1939) suggested even more restricted conditions, and postulated that gilgaies develop only from alkaline solonetz soils, in which, after heavy rains, there is differential absorption at certain points when water from the saturated A horizon penetrates the otherwise impermeable B horizon around root channels and burrows, causing isolated pressure zones which force blocks of soil upwards.

From general considerations of the distribution, character, and size of the gilgaies of New South Wales, it can be suggested that there are two sets of factors related to their formation. These are the external factors of the site—rainfall and temperature fluctuations, likelihood of flooding by rain falling higher up the catchment area, the frequency of droughts, &c.—and the internal factors due to the character of the soil material, such as those discussed here. Neither of these two sets of factors need have remained constant during the period of soil formation, and it is not improbable that in parts of Australia formations will be found that can only be interpreted as fossil gilgaies, formed perhaps during the Arid period of Recent time. No such fossils have been found by the authors, and all gilgaies studied appear to be either still forming or at least self-maintaining at the present time.

If it can be postulated that maximum development of the gilgaies is to be expected when one or more of the factors responsible is at its maximum, it should be possible to find a relationship between these factors and the amplitude of the gilgai formation at each site. Both external factors and internal properties are concerned, for the amplitude must be related in some way to both the extent and frequency of the movement. Of the external factors, the frequency and extent of wetting and drying are clearly important, but this aspect will be discussed later.

Restricting consideration to the internal properties, if the undulations of the surface are the result of soil movement their amplitude must be related to the ability of the soil both to develop a thrust and to transmit it. The ability to develop a thrust is measured by the volume expansion, and from the figures presented here the swelling profile is quite characteristic. In gilgaies the volume expansions of the topsoils are not less than 10 per cent., whilst those of the deep subsoils are considerably in excess of the topsoils. These changes are associated with increases down the profile of clay content, exchange capacity, and the proportion of sodium ions on the exchange complex. At the same time the whole solum contains a higher proportion of exchangeable sodium and a higher sand content in the gilgaied soils than in the non-gilgaied variants of the same group, whilst on the other hand it has already been shown (Fig. 2) that there is a crude inverse relationship between clay content and vertical

interval of the gilgai. These features are probably related to the ability of the soil to transmit a thrust. In any soil the particles are always aggregated to some extent, and elementary considerations suggest that the size of the aggregate will have a bearing on the extent of the deformation of the surface. Thus a thrust applied to a mass of small aggregates will be diffused through the system in a manner somewhat

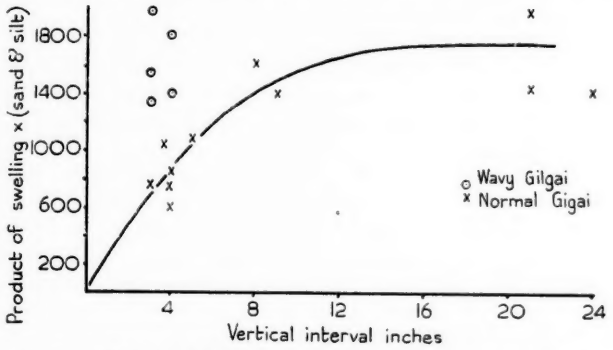


FIG. 7. Relationship between a parameter of structure (viz. swelling × [sand + silt]) and the vertical interval for some normal and wavy gilgaies.

analogous to the diffusion of a thrust through a fluid. As the aggregates increase in size, the effect of the thrust becomes progressively less dispersed, whilst with large aggregates a thrust produces a definite movement of one block of the soil relative to the other blocks on either side of it. From this the vertical interval between the undulations of the surface might well be related to the ability of the soil to form large aggregates.

Attempts were consequently made to find some parameter, incorporating both swelling and mechanical composition, which could be related to the vertical interval. Briefly, such relationships were found with all parameters that were products of volume expansion and some component of the non-clay fraction, and for one of them the relationship is shown graphically in Fig. 7. No similar relationships could be found to the wave-length.

It is clear from this that whilst a curvilinear relationship holds between this parameter and the vertical interval for normal gilgaies, no such relationship holds for wavy gilgaies. Since wavy gilgaies are always formed on sloping ground, whereas normal gilgaies are found on the flat, the amplitude of development in wavy gilgaies may be effected by processes of erosion as well as by the internal properties of the soil.

Restricting further discussion to normal gilgaies, it can be seen that the vertical interval between puff and shelf is related to both the swelling capacity of the clay* and the percentage sodium saturation of the

* Obtained by dividing the percentage expansion by the percentage clay of the sample concerned.

exchange complex (Figs. 8 *a* and *b*), and the relationship in each case is significant.

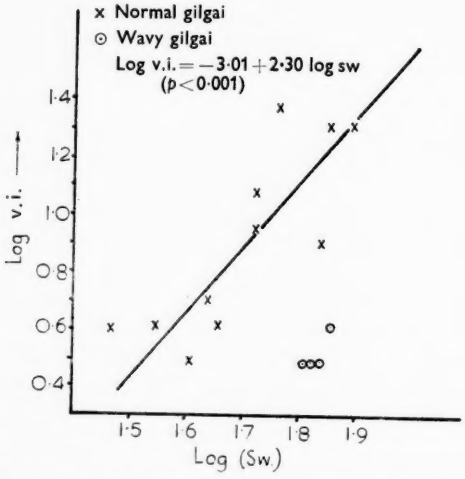


FIG. 8 *a*. Relationships between vertical interval and swelling capacity.

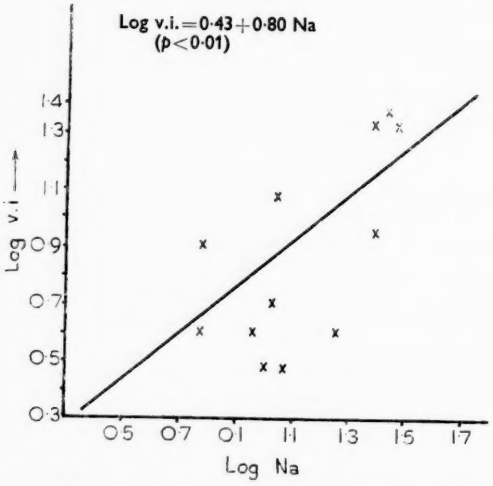


FIG. 8 *b*. Relationship between vertical interval and percentage sodium saturation of the exchange complex.

Since the swelling capacity of the clay is partly dependent on the degree of saturation by sodium, the effect of sodium could arise solely from its effect on the swelling capacity. Regression analysis of the

relationship between vertical interval, swelling capacity, and per cent. sodium saturation gives the equation:

$$\log v.i. = 2{\cdot}71 + 1{\cdot}72 \log sw + 0{\cdot}38 \log \mathrm{Na},$$

where $v.i.$ = vertical interval in inches,

sw = swelling capacity of the clay,

Na = percentage saturation of the exchange complex by sodium.

For sw, $t = 3{\cdot}23$ ($P < 0{\cdot}01$) and $r_{sw} = 0{\cdot}835$ ($P < 0{\cdot}001$).

For Na, $t = 1{\cdot}91$ ($P < 0{\cdot}1$) and $r_{\mathrm{Na}} = 0{\cdot}732$ ($P < 0{\cdot}01$).

From which it follows that the swelling capacity of the clay is significantly related to the vertical interval of the gilgai, whilst the degree of sodium saturation of the exchange complex has a significant effect, which is independent of its effect on the swelling capacity of the clay.

Expressed in another way, 53 per cent. of the variation in vertical interval is due to variation in swelling capacity of the clay, and 28 per cent. due to an effect of exchangeable sodium other than its effect on swelling, leaving 19 per cent. of the variation as due to all other factors and error.

It has already been shown (E. W. Russell, 1934) that the sodium ion produces larger and more resistant clods, and it seems likely therefore that the effect of sodium saturation *per se* upon the vertical interval is directly due to this effect, and the larger the clod moved by the swelling of the subsoil the greater would be the amplitude of the undulations produced.

From the foregoing account it becomes possible to offer an explanation of the differences between gilgai soils and their distribution. In the first place it is clear from the field descriptions and from the analytical figures that the puff is largely made up of subsoil that has been forced to the surface, whilst in the second place, the extent to which the soil of the gilgaies swells when wetted indicates the source of the thrust necessary to produce this movement.

If the shelf profile is considered as the more or less normal soil immediately before movement commences, the mechanism would appear to be as follows:

As the soil dries out, the topsoil dries first, shrinking and cracking as drying proceeds. This cracking aids the drying out of the deeper layers, which, as drying continues, also shrink, ultimately to a considerably greater extent than the topsoil (indicated by their greater swelling capacity, and illustrated in Plate II *b*). When re-wetting takes place, either from floodings or from rain, the whole soil swells again. If the drying out and subsequent shrinkage had been uniform, and had decreased uniformly from surface to subsoil, re-wetting and consequent swelling would merely have restored the soil level to what it was before. In a deeply cracking soil, however, uniformity of drying and shrinkage is unlikely, since the moisture content will show a horizontal gradient from the face of each crack into the interior of the block, as well as the general vertical gradient from surface to subsoil. Re-wetting will be also uneven. Light showers will merely wet the surface, whilst heavy rains, particularly

the summer rains which are not uncommonly torrential in character, will wet and swell the subsoil at the base of the cracks long before the rest of the subsoil at the same level has had any change in moisture content. Uneven re-wetting is consequently likely to be the usual occurrence on such soils.

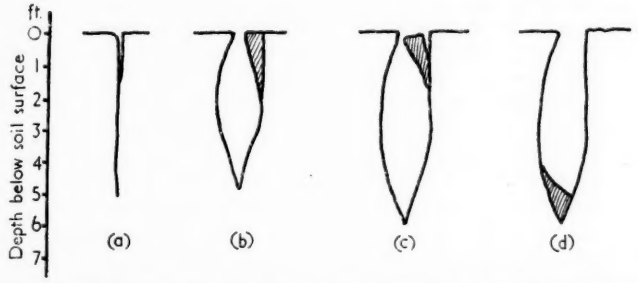
The cracks themselves also contribute indirectly to the uneven distribution of pressures on re-wetting. During the dry period pieces of surface soil fall down the cracks, and the fact that the subsoil shrinks more than the topsoil, giving cracks that are wider below than they are at the surface, makes it easy for large pieces of the surface soil to break off and to slip down into the crack below. Subsequent re-wetting and re-hydration would lead to excess pressure causing the fracturing of blocks of soil from that lying below (Fig. 9). Repetition of this process would cause the blocks to move upwards towards the surface, since subsequent cracking and fracturing would start to occur again about the position of the earlier cracks, which would have become lines of weakness. As the blocks move upwards rain and wind would erode the surface soil and distribute it over the surrounding soil surface, so that in time the blocks that had previously been subsoil would be exposed at the top. This explanation, advanced originally by Leeper (1936), is supported by all the field descriptions and accords with the swelling profile of the gilgai soils given in Fig. 6.

The different forms of gilgai encountered, and their absence from certain soils, follow from changes in the nature of the profile or changes of the environment. Some of these points may be illustrated by reference to the soils whose swelling profiles are illustrated in Fig. 6.

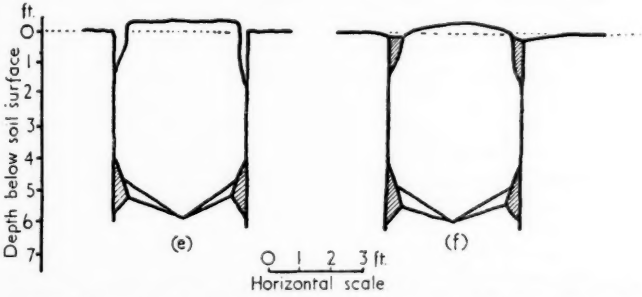
If the surface soil is sandy and does not crack, more uniform drying of the subsoil will obtain. At the same time the dry layer formed acts as a surface mulch and lowers the temperature and hence the rate of evaporation of water from the deep subsoil. The frequency as well as the magnitude of the volume changes of the subsoil will consequently be at a minimum, and in the extreme case no gilgaies will form, even in soils such as the solodized-solonetz or the solodic soils, in which the swelling profiles show a qualitative resemblance to those of the gilgai soils. By the same token, gilgaies would not be expected where for any other reason the volume changes in the subsoil would be at a minimum. This would occur where the subsoil lacks the ability to swell, as in the sandy podzols of the coast or any other soils with a sandy subsoil. It would also hold where the profile is too shallow, as in the semi-skeletal soils of the hill crests and ranges, as for example in the stony chocolate soils of the Namoi region. The swelling profile of the deeper chocolate soils also show qualitative similarities to those of the gilgai soils, but the absence of gilgaies on these soils is explained by the quantitatively lower order of the swelling capacities, and by the more equable climate of those areas where they show their optimum development (Hallsworth, Costin, Gibbons, and Robertson, 1952).

The absence of the gilgai phenomenon from other, and commonly adjacent, heavy soils of the sub-humid and semi-arid regions of the State also fits this pattern of soil movement. It can be seen from Fig. 6 that

for the topsoils of the normal chernozemic and sierozem soils the swelling is of the same order as it is for the subsoils. Since the subsoil would not dry out to the same extent, nor with the same frequency as the topsoils, the movement of subsoil relative to the topsoil would be less



The development of soil cracks, drawn to scale using the swelling figures to the Gilgandra gilgai (Profile Nos. 51/52), and assuming that the primary cracks formed at 50 in. intervals. (*a*) Primary crack developing. (*b*) Soil completely dried out and showing maximum shrinkage to 3 ft.; still moist and showing no shrinkage at 5 ft.; with a secondary crack developing at the side. (*c*) Soil completely dried out and showing maximum shrinkage to 4 ft.; the moist zone has retreated to 6 ft.; the secondary crack has extended, causing the fracture of a piece of surface soil, which falls to the bottom (*d*).



Effect of complete re-hydration after heavy rain or flooding, assuming for simplicity that all strain has been taken by the central block. (*e*) Surface material at bottom of crack leads to fracture of the block from underlying soil, and forces a roughly equal volume of soil above the surface. (*f*) The holes left by the fallen blocks are filled up by the in-washing of surface soil from either side of the crack. Subsequent drying and re-wetting will lead to a repetition of the process.

FIG. 9. Diagrammatic representation of the formation of a gilgai.

and the chances of pieces of topsoil falling into the subsoil would be minimized, and the likelihood of soil heaving due to unequal stress is consequently reduced.

This concept also relates the variable distribution of the gilgaies across the great alluvial plains to the character of the alluvium. Where the latest deposit has been clayey in nature and overlies a lighter textured

subsoil, no gilgaies will develop, and normal chernozemic or sierozem soil will be found. Where the latest deposit has been of sandy material showing literally no shrinkage, then no gilgaies will develop whatever the nature of the underlying deposit. If, however, the underlying deposit has a fairly high swelling capacity, and if the latest alluvium contains sufficient clay to allow appreciable shrinkage to occur, gilgaies will develop. In the same way the solonization of chernozem or sierozem soil, with the migration of clay to lower horizons and the increase in the exchangeable sodium, would lead to the development of the typical swelling profile, and subsequent development of the gilgai phenomena.

The difference in the character of the puff and shelf follows from the character of the soil material. Soils in which large aggregates are formed will have large blocks moving into the surface, and the longer they remain coherent the larger will be the size of the 'puff' and the greater the vertical interval (viz. the Gilgandra profile, Nos. 51 and 52). Where the subsoil moving up to the surface contains large quantities of calcium carbonate or gypsum, the blocks will tend to disaggregate, and fall down into much smaller granules as soon as they slake under the influence of rain, giving the characteristic 'puffy' structure. The resulting mound will then have more gentle contours (as in the Pallamallawa profile, No. 160 *a* and *b*). In general, therefore, the strongly calcareous soils will not show as large a vertical interval as the less calcareous ones, other conditions being constant.

The sandy and sodium-saturated soils of the stony gilgaies show the opposite feature of a tight and compact 'puff', but only show a small vertical interval. Since the vertical interval is obviously determined by the frequency as well as the magnitude of the volume changes, and climatic records suggest that only once in every 25 years will the rainfall on the Stony Downs be enough to saturate the subsoil, it is obvious that the volume changes will occur only at very infrequent periods and the small development of the vertical interval is consequently due to limitations of climate rather than of the soil material.

The wavy gilgai are found only on sloping ground and on a more restricted range of parent materials. Those seen by the authors have been invariably calcareous, and with a very granular puff. The combination of the granular nature of the protruded soil and the slope of the soil surface leads to the ready erosion of those parts of the 'puff' which run across the line of slope, and at the same time preserving the other parts of the puff. Repetition of the process of soil heave and surface erosion will then produce the gentle curves of the wavy gilgai, running more or less up and down the line of slope. It might be added here that quite normal stony gilgai have been observed north of Broken Hill lying on slopes which in the central regions of the State would have carried wavy gilgai, and it would seem that the very heavy stone layer acts as a mulch resisting the effects of erosion.

The melon-hole gilgai show certain features of difference from the other gilgai soils. In the first place they are commonly very acid and completely lacking in calcium carbonate. Since they are still forming at the present time, it is clearly not possible to accept Howard's limitation

of gilgai formation to the alkaline solonetzic soils. The low calcium status is reflected in the character of the mounds, which are so massive in structure as rather to belie the term 'puff'. Apart from the small mounds mentioned earlier, one is more conscious of the process of soil sinking down into holes rather than being forced up into the puffs. Thirdly they occur in much higher rainfalls than the other forms, and it can only be very rarely indeed that the subsoil dries out thoroughly. However, the very high clay content of these soils allows considerable changes in volume even without complete drying out, and the drier lumps of surface soil might be expected to re-hydrate after slipping down into the cracks of the still relatively moist subsoil, and to be merged into it with less strain than would occur in the drier areas of the chernozemic and sierozem soils.

One other point may be important here. Much of this melon-hole country was originally levelled for sugar-growing, but the heavy texture rendered cultivation uneconomic, and other parts have since been levelled for the development of sown pastures. Melon-holes re-form over all these areas, and quite rapidly after a flood. It seems probable that the small mounds of soil found on the 'puffs' of the recently redeveloped gilgaies could be due to semi-fluid soil having been forced up between cracks of the clay during the very rapid and vigorous swelling of the subsoil during the flood. This form of movement is probably very subsidiary (in quantity) to the movement of blocks by fracture, but may be important in areas liable to flooding, either from river water or by very heavy rain. The small 'flows' of subsoil observed at Gilgandra would also fall into this class of movement.

The climatic control of the distribution of gilgaies follows from its control of the frequency of soil movement. Obviously there will be no soil movement if the subsoil remains permanently wet and hydrated to its fullest extent. At the other extreme there will be no movement if the subsoil remains perfectly dry. In between these extremes soil movement will take place, and will reach its maximum where the maximum change of moisture content in the subsoil occurs most frequently. In New South Wales this zone of maximum movement appears to lie between the rainfall limits of 14 to 24 in. in the north and 12 to 22 in. in the south. With higher rainfalls, the subsoil will dry out less frequently, and shrinkage will not be so complete. At lower rainfalls, as in the Western Division, the subsoil will be wet less frequently, and from either cause the movement will be more restricted.

The occurrence of the melon-hole gilgai under the higher rainfalls of the north coast is explained by the much higher temperatures and by the occasional droughts. The climate of the north coast appears to represent the limit under which gilgaies will form, for they are here restricted only to the very heaviest of the soils, with clay contents ranging between 70 and 90 per cent.

The absence of gilgaies from the tablelands is attributable to the higher effective rainfalls and the lower summer temperatures, which would limit the drying out of the subsoil. Their absence from the extensive areas of the lower rainfall region of the Monaro appears to be due to the

much lower soil temperatures, whilst such soils as have been analysed have shown also a lower level of exchangeable sodium in the subsoil.

The frequency with which gilgaies have been observed in Australia compared to the infrequency of similar reports elsewhere is a matter for comment. They have been observed in Texas (Stephens, 1948) and in the Orange Free State (Voster, 1950, priv. comm.), whilst Milne (1942) describes a patchy distribution of calcium carbonate nodules on the surface of a soil in Tanganyika which strongly resembles some of the occurrences in New South Wales. Nevertheless, it is clear that neither in magnitude nor extent are gilgaies as numerous elsewhere as they are in Australia. This is probably due to the combination in Australia of a climate providing optimum conditions for gilgai formation with large areas of soils of suitable physical and chemical composition.

Summary

A field and laboratory study has been made of a common feature of the soils of sub-humid and semi-arid areas of Australia, the 'gilgaies', in which undulations of the surface of the land occur, the alternate hummocks and hollows of which commonly show some degree of regularity.

Six major groups have been recognized in the field, which have been termed (1) round or normal gilgai, (2) lattice gilgai, (3) wavy gilgai, (4) tank gilgai, (5) stony gilgai, and (6) melon-hole gilgai. The vertical interval observed varies from ½ in. to 10 ft. and the wave-length from 4 ft. to 100 ft., and the phenomenon is found as a variant of meadow, chernozemic, sierozem, and stony downs (stony desert) soils, under rainfalls ranging from 6 to 60 in.

An examination has been made of their chemical character, mechanical composition, and clay mineralogy, as well as the climatic and topographic limitations to their development in the field.

It has been found that the distinguishing feature is that all gilgai soils show a characteristic swelling pattern on wetting, the topsoil possessing a certain minimum swelling capacity, which is increased considerably in the subsoil. The swelling pattern of the non-gilgaied soils can be divided into two groups: either the swelling capacity of the subsoil is similar to or less than the topsoil, or, where an increase in swelling capacity takes place in the subsoil, the swelling of the topsoil is less than 10 per cent.

The swelling of the soil has been shown by statistical analysis to be significantly related to the clay content, the exchange capacity of the clay, and the amount of sodium in the exchange complex, and an increase with depth of these three factors has been demonstrated on the shelves of all the gilgai soils examined, irrespective of the locality or soil type on which they have been formed.

An investigation has also been made of the relationships between the internal properties of the soil and the amplitude of the phenomenon, and it has been found that, in the zone of optimum gilgai development, the vertical interval between puff and shelf is significantly related to the swelling capacity of the clay and to the percentage sodium saturation of the exchange complex.

Acknowledgements

The authors are greatly indebted to the Research Grants Committee of the University of Sydney for their generous financial assistance which made this work possible, to the Rural Bank of New South Wales for financial assistance in the field work, to Miss H. V. Raynor for certain analyses, and to their colleagues in the University, the New South Wales Department of Agriculture, the Forestry Commission, and the Water Conservation and Irrigation Commission for their constant interest and assistance, and in particular to those former students who supplied the photographs included here.

REFERENCES

Black, C. A. 1947. Soil Sci. **3**, 337.
Coutts, J. R. H. 1930. J. Agric. Sci. **20**, 407.
Doughty, J. L. 1941. Sci. Agric. **22**, 135.
Ellis, J. H., and Shafer, W. 1928. Ibid. **9**, 231.
Hallsworth, E. G., Colwell, J. D., and Gibbons, F. R. 1953, Aust. J. Agric. Res. **4**, 305.
—— Costin, A. B., Gibbons, F. R., and Robertson, G. K. 1952. J. Soil Sci. **3**, 103.
—— —— —— 1953. Ibid. **4**, 241.
Howard, A. 1939. Pedology No. 8, 14.
Krishna, P. G. 1950. Soil Sci. **70**, 375.
—— and Perumal, S. 1948. Ibid. **66**, 29.
Leeper, G. W., Nicholls, Ann, and Wadham, S. M., 1936. Proc. Roy. Soc. Victoria, **49**, 77.
McElroy, C. T. 1952. J. Roy. Soc. N.S.W. **85**, 53.
MacEwan, D. M. C. 1946. J. Soc. Chem. Indust. **65**, 298.
Milne, G. 1947. J. Ecol. **35**, 192.
Nikiforoff, C. C. 1941. U.S.D.A. Tech. Bull. No. 745.
Prescott, J. A., 1932. C.S.I.R. (Aust.) Bull. No. 52.
Russell, E. W. 1934. Phil. Trans. **233** A, 361.
Sigmond, A. A. J., de. 1937. The Principles of Soil Science. Thos. Murby, London.
Stephens, C. G., 1948. Overseas Memoranda to Members of Division of Soils C.S.I.R. and other Australian Soil Scientists. (Mimeo.)

(*Received 14 September 1953*)

A NOTE ON GILGAIES AND FROST SOILS

A. B. COSTIN*
(*Soil Conservation Authority of Victoria*)

LARGE areas of Australia are characterized by gilgai soils (cf. Hallsworth, Robertson, and Gibbons, 1955). At high latitudes (e.g. Lappland) and to a less extent in alpine environments, deep-frost soils are equally conspicuous (cf. Lundqvist, 1948). Yet the striking similarity between the surface features of the soils in each case has brought forth surprisingly little comment; this, no doubt, is due to the impression that since the respective environments are so different the soils can have little in common either. However, we admit the limitations of the zonality or monoclimax theory for soils and vegetation, and recognize the fact that the same soil or plant community may be zonal or climatic in one area but limited by some factor other than climate elsewhere. We must likewise concede that similar soil properties may arise by different soil-forming processes. Considered in this light, the similarities between the surface features of gilgai and deep-frost soils may be examined further.

Six types of gilgai have been recognized in New South Wales (Hallsworth *et al.*, 1955); for the present purposes, however, these may be reduced to three on the basis of their approximate relation to slope: round or normal gilgai, lattice gilgai, and wavy gilgai. In the normal gilgaies, usually occurring on level ground, the alternating hummocks (puffs) and hollows (shelves) are subcircular in outline. On very slight slopes there is a gradation from the normal type to the lattice form in which either the puffs or shelves are discontinuous and orientated in the direction of slope, producing a lattice or network pattern. The wavy type usually develops on slightly greater slopes and takes the form of continuous puffs and continuous shelves arranged in regular lines more or less parallel to the direction of greatest slope or in rare cases (not properly understood) transversely to it.

Corresponding surface features are developed in the deep-frost soils. On level ground these features may be expressed as alternating hollows and hummocks (Plate I†) or large stone rings or stone nets of subcircular or polygonal shape. On slight slopes these normal forms first grade into a crescentic type (Plate II *a*†) in which the successive crescents sometimes have the appearance of broad steps, then into a wavy type in the form of long, parallel undulations running regularly down or across the slope (Plate II *b*†), or large stone stripes (Plate III *c*†) or rock rivers (Plate III *b*†). The crescentic form of deep-frost soil appears to differ somewhat from the corresponding lattice gilgai in its lack of distinct downslope orientation.

* The writer is indebted to Mr. F. R. Gibbons for his criticism of the manuscript.

† These plates may be compared with those illustrating gilgai features in the paper by Hallsworth *et al.*, 1955.

This parallelism between the gilgai and deep-frost soils is so striking in the field that one immediately wonders if the processes responsible in each case have anything in common.

Although in New South Wales gilgaies occur on several soil groups and over a wide range of climate (e.g. average annual rainfall from 7 to 70 in.), their formation has been shown to require certain well-defined edaphic and climatic conditions (Hallsworth *et al.*, 1955). The subsoil clays of gilgai soils possess a swelling capacity of at least 30 per cent. and have sufficient exchangeable sodium in the B_2 horizon to induce a coarse cracking pattern; sufficient seasonal difference (in rainfall and/or temperature) is also necessary to allow the cracking properties of the soil to express themselves. This combination of conditions results in sufficient differential wetting and drying of the subsoil and hence sufficient differential swelling and shrinking to allow the production of isolated pressure zones with the consequent development of gilgai features. Operating on approximately level ground, the normal form of gilgai is thus produced; when the effects of slope are superimposed the lattice and wavy types are formed. In all cases the differential growth of vegetation may operate as a subsidiary influence and accentuate the gilgai features.

For the development of deep-frost soils a suitable alternation of climate and a suitable subsoil condition are also necessary. In this instance, of course, the swelling and shrinking of the subsoil are achieved not by alternate wetting and drying but by freezing and thawing, and the physical condition of the subsoil is such that the expansion thrusts are differentially applied so as to produce differential movement. The normal forms of deep-frost soil are produced where these conditions operate on approximately level ground, and the crescentic and wavy forms where the effects of slope are superimposed. As with gilgaies, the differential growth of vegetation may accentuate these features.

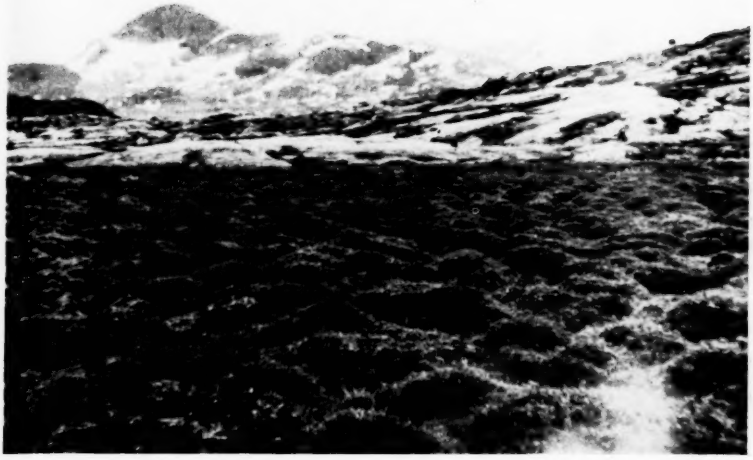
So far subsoil swelling and deep-frost effects have been considered. Let us now examine purely surface phenomena. The polygonal surface-cracking pattern (often combined with small surface undulations) which is developed by soils capable of swelling and shrinking on wetting and drying is almost too common to need mention. The corresponding surface-freeze phenomenon on level ground is the small stone ring or stone net, which on sloping ground grades into small stone stripes; if the soil lacks coarse particles or is of uniform particle-size, however, characteristic frost-fluted surfaces (resembling small, irregular corrugations) are developed instead (Costin, 1950). On slopes saturated by melt-waters from adjoining snow patches (where surface freezing and thawing regularly occur during the snow-free months) small solifluxion terraces are formed (cf. Watt and Jones, 1948); these are the counterparts of the crescentic deep-frost soils and of the lattice gilgaies.

These brief notes on the similarities between the surface features of gilgai and deep-frost soils and between surface-cracking and surface-freeze phenomena serve to illustrate the important principle of 'factor compensation' in pedology and ecology (Costin, 1953). This principle states that similar soils and soil properties and similar plant communities may be produced from different combinations of environmental factors.

REFERENCES

COSTIN, A. B. 1950. J. Soil Cons. Serv. N.S.W. **6**, 73.
—— 1953. A Study of the Ecosystems of the Monaro Region of New South Wales. Govt. Printer, Sydney.
HALLSWORTH, E. G., ROBERTSON, G. K., and GIBBONS, F. R. 1955. J. Soil Sci. **6**, 1.
LUNDQVIST, G. 1948. De svenska fjällens natur. Svenska Turistföreningens Förlag, Stockholm.
WATT, A. S., and JONES, E. W. 1948. J. Ecol. **36**, 283.

(*Received 1 January 1954*)



a. Normal type or deep-frost soil: the analogue of the normal gilgai. (Near Riksgränsen, Swedish Lappland, 6 52)

b. Transitional stage between the normal and crescentic types of deep-frost soil. (Near Abisko, Swedish Lappland, 6 52)

A. B. COSTIN—PLATE I

a. Crescentic type of deep-frost soil, in this case resembling broad steps: the analogue of the lattice gilgai. (Near Abisko, 6/52)

b. Wavy type of deep-frost soil: the analogue of wavy gilgai. (Near Abisko, 6/52)

A. B. COSTIN—PLATE II

a. Giant stone stripes: a variant form of the wavy deep-frost soils. (Near Narvik, Norwegian Lappland, 7/52)

b. Rock-river (left foreground). (Near Abisko, 6/52)

A. B. COSTIN—PLATE III

ALPINE SOILS IN AUSTRALIA

WITH REFERENCE TO CONDITIONS IN EUROPE AND NEW ZEALAND

A. B. COSTIN

(*Soil Conservation Authority of Victoria*)

Introduction

THE soils of alpine areas in general are among the least studied of all soil groups. Apart from some detailed information on the soils of the Continental Alps and the alpine humus soils of New South Wales (Costin, Hallsworth, and Woof, 1952; Costin, 1954) virtually no comprehensive surveys appear to have been reported in the literature. Much of the early information published on the Australian alpine soils, furthermore, gives an erroneous impression that they are predominantly skeletal, shallow, and immature.

This dearth of information on alpine soils is accentuated by the inconsistent usage of the term 'alpine'. Some workers would regard as alpine all areas beyond the winter snowline; others would restrict the term to the very steep, snow-free areas within the tract of permanent snow. The most useful and natural usage is intermediate between these extremes: it regards as alpine *elevated environments above the tree-line where the ground is snow-covered for about six to twelve months of the year* (cf. Beadle and Costin, 1952). Even when this definition is adhered to, however, the alpine group of soils as a whole includes many heterogeneous types, such as the 'frost-gilgaied' saline soils of dry arctic-alpine regions, rendzinas and brown limestone soils in the European Alps, bog peats in more oceanic European mountains, and alpine humus soils in Australia; in one and the same alpine area, moreover, the soils may range from acid peats to alkaline mineral types. To refer to soils as alpine, therefore, may give no information on their characteristics; at the most the term can be applied to a complex, or an association of soils, the members of which should be referred to individually.

The Alpine Environment in Australia

The only truly alpine environment of the Australian mainland occurs in the Snowy Mountains of New South Wales, with smaller outliers in the Victorian Alps. It extends from the natural tree-line of *Eucalyptus niphophila* (snow gum) at about 6,000 ft. to the maximum elevation of 7,328 ft. on Mount Kosciusko. These alpine areas represent the most uplifted portion of the old Tertiary peneplain, now much dissected by post-Kosciusko erosion. They differ notably from the alps of most overseas fold mountains in their comparative lack of precipitous topography. These gentler slopes, combined with a less severe Pleistocene history, have favoured soil development to an unusual degree. In fact, it might be said that the Australian Alps are in the nature of soil mountains, in

contrast to most others which are more predominantly rock, as in Switzerland and New Zealand, or peat as in the oceanic mountains of Europe. A further implication of these gentler slopes is the continuity of the Australian alpine environment compared with the marked discontinuity which is such a feature of the steeper mountains elsewhere. Indeed, it is doubtful if a more extensive and continuous low-alpine tract than that of the Snowy Mountains exists anywhere in Europe.

The major rock types of the Australian Alps are gneisses and granites, with smaller occurrences of slate, similar metamorphosed rocks, and occasionally basalt.

Some aspects of the alpine climate on the Australian mainland may now be mentioned. The prevailing south-westerly to north-westerly winds bring most of the precipitation which falls with a strong relative winter incidence and occurs mainly as snow. These winds may at times blow with great force: at exposed alpine levels at Kosciusko, for example, saltation movements involving gravel of ½-in. diameter are common. The amount of precipitation varies from about 70 to 90 in. per annum and the number of rain days is probably in the vicinity of 130 to 150. Meyer ratios vary from approximately 1,000 to 1,500. The mean monthly maximum temperatures rise to about 55° to 60° F. in January and February, but approach freezing in midwinter; the mean monthly minima fall to about 20° F. in the coldest month and do not exceed freezing for about 5 to 8 months of the year. Whereas only small daily temperature fluctuations are experienced during the period of winter snow cover, the daily range may at times approach 50° F. in fine summer weather when loss and gain of heat are rapid. During the snow-free months (especially in spring and autumn) severe ground frosts may result in a regular daily freezing and thawing of the topsoil; in this daily temperature environment the alpine soils of Australia differ from those of most homologous low-alpine areas of Europe. In winter, however, it is doubtful if the meteorological data available are of much pedological significance, since the snow-cover insulates the soils from atmospheric conditions to the extent that they rarely freeze.

The climax alpine plant communities* include tall and short alpine herbfields, sod tussock grassland, fen, raised and valley bogs, fjaeldmark, and heath (Table 1). The extensive tall alpine herbfields of the *Celmisia longifolia–Poa caespitosa* alliance constitute the climatic climax, being replaced at subalpine levels by subalpine woodlands of the *Eucalyptus niphophila* alliance, and under special physiographic and edaphic conditions in the alpine tract by the other communities mentioned in the Table.

The primary biological complex—i.e. the flora and fauna as distinct from the dependent plant and animal communities—seems to be of unusual importance in the genesis of Australian alpine soils. The outstanding species in this respect are the grass *Poa caespitosa* and the mountain earthworm *Megascolex* sp.; the former appears not only to circulate soil bases but to mobilize and concentrate soil sesquioxides near

* The classification and nomenclature of the vegetation is in accordance with the scheme proposed by Beadle and Costin (1952).

the surface, and the latter, by virtue of its vigorous development under strongly acid conditions, may contribute still further to these effects (cf. Costin, Hallsworth, and Woof, 1952).

More recently the human or anthropogenic factor has assumed importance in the alpine environment. Summer grazing of sheep and cattle, with the associated practice of burning, has been responsible for considerable deterioration of the soils and vegetation. In some areas of the Snowy Mountains this adverse trend is now being accelerated by maladjusted engineering operations.

The Alpine Soils of Australia

Far from being skeletal, shallow, and immature, the alpine soils of the Australian mainland are for the most part well developed and can be readily differentiated: of the soil groups recognized so far in Australia by Hallsworth and Costin (1950), as many as eight great soil groups with six subgroups occur in the alpine tract. A classification of these alpine soils is included in Table 1.

As a preliminary to the description of individual alpine soils, relevant environmental data are also tabulated. It may also be mentioned that only one of these groups, namely the snow patch meadow soils, is exclusively alpine in Australia, and only the alpine humus soils are virtually alpine-subalpine in distribution. Whilst some of the remaining groups undergo optimum development in alpine-subalpine environments, they also occur at lower levels.

Descriptions of individual profiles and analytical data are given as an Appendix.

Alpine humus soils

The alpine humus soils of New South Wales have been considered in some detail elsewhere (Costin, Hallsworth, and Woof, 1952). Their status may be regarded as that of a climatic climax, in so far as climate is the factor which most limits their development.

Lithosols

In normal alpine situations the ground is protected from the sudden atmospheric temperature changes of late autumn, winter, and early spring by a cover of snow, under which conditions alpine humus soils with closed tall alpine herbfield usually develop. On the other hand, many steep upper slopes and exposed peaks and plateaux on which snow is prevented from accumulating for so long are not so effectively protected; similar exposure effects also occur in summer near snow patches where freezing and thawing are commonly daily phenomena. It is in these steep, exposed, and snow patch situations where wide daily temperature fluctuations with frost and ice action are experienced that the alpine lithosols are best developed (cf. Costin, 1950); their status is clearly that of a physiographic climax.

In some areas of the alpine tract, however, lithosols are also found in the relatively protected situations apparently favourable for alpine humus

TABLE 1

Summarized Environmental Data on the Alpine Soils of the Australian Mainland

Great Soil Group	*Subgroups*	*Physiography*	*Geology*	*Vegetation*	*Status*
Soils in which the profile shows no eluviation of sesquioxides. Acid to strongly acid in the surface					
ALPINE HUMUS SOILS . .		Wide range of physiography, except where extremely windswept or impeded drainage.	Independent of rock type.	Mainly tall alpine herbfield (*Celmisia longifolia–Poa caespitosa* alliance) and sod tussock grassland (*Poa caespitosa–Danthonia nudiflora* alliance), with some fjaeldmark (*Epacris petrophila–Veronica densifolia* alliance) and heath (*Oxylobium ellipticum–Podocarpus alpinus* alliance).	Climatic climax.
Soils showing no profile differentiation. Largely mineral in character					
LITHOSOLS .		Steep upper slopes, snow patches, and exposed peaks and plateaux.	Wide range of rock types, usually strongly jointed or fractured	Mainly fjaeldmark (*Epacris petrophila–Veronica densifolia* and *Coprosma pumila–Colobanthus benthamianus* alliances) and heath (*Oxylobium ellipticum–Podocarpus alpinus* alliance).	Physiographic climax.
Soils in which a water table is present in the solum for at least part of the year					
GLEY PODZOLS .	Sidero-gley Podzol. Organo-gley Podzol.	Level to moderately sloping situations, influenced by an acid water table in the subsoil.	Usually alluvial and colluvial parent materials, derived from most rocks.	Sod tussock grassland (*Poa caespitosa–Danthonia nudiflora* alliance) and heath (*Epacris serpyllifolia–Kunzea muelleri* alliance).	Physiographic climax.
SILTY BOG SOILS		Level to gently sloping situations, influenced by a permanent, acid water table near surface.	Alluvial and colluvial parent materials, derived from a wide range of rock types.	Sod tussock grassland (*Poa caespitosa–Danthonia nudiflora* alliance) and fen (*Carex gaudichaudiana* alliance).	Physiographic climax.
MEADOW SOILS .	Snow Patch Meadow Soil.	Considerable range of slopes, but confined to base of semi-permanent snow patches on leeward aspects.	Fine-textured parent materials derived from a wide range of rock types.	Short alpine herbfield (*Plantago muelleri–Montia australasica* alliance).	Physiographic climax.

TABLE 1 (*cont.*)

Great Soil Group	*Subgroups*	*Physiography*	*Geology*	*Vegetation*	*Status*
Soils in which the profile is dominated by organic matter, of which the botanical features are partially visible					
BOG PEATS	Valley Bog Peat.	Level to gently undulating situations, permanently wet.	Independent of rock type. Acid and/or base deficient water conditions.	Valley bog (*Carex gaudichaudiana–Sphagnum cymbifolium* alliance).	Physiographic climax.
	Raised Bog Peat.	Sloping situations, permanently wet.	Independent of rock type. Slightly more acid and/or base-deficient water conditions than the valley bog peats.	Raised bog (*Epacris paludosa–Sphagnum cymbifolium* alliance).	Physiographic climax.
FEN PEATS	Poor Fen Peat (eq. Acid Lowmoor Peat).	Level to gently sloping situations, permanently wet.	Largely independent of rock type. Waters acid but with a small base supply.	Mainly fen (*Carex gaudichaudiana* alliance), occasionally short alpine herbfield (*Plantago muelleri–Montia australasica* alliance).	Physiographic climax.
HUMIFIED PEATS		Considerable range of slopes.	Derived largely from deposits of bog peats.	Tall alpine herbfield (*Celmisia longifolia–Poa caespitosa* alliance).	Climatic climax where formed as result of natural drainage; anthropogenic climax where drainage has been caused by human interference.

pedogenesis. These appear to be of considerable palaeo-pedological significance, in so far as they have features in common with the 'rock rivers' and similar periglacial soils now forming in Lappland and even there also regarded in some parts as relics of more severe Pleistocene conditions.

The lithosol profile typically shows no horizon differentiation; it consists largely of angular rock fragments which may vary in size from large stones to gravel, depending on the type of rock and conditions of physical weathering. Two main types may be recognized: the first consists for the most part of large stones, sometimes to a depth of several feet, formed by sudden temperature changes and other resultant processes

of frost action and nivation; the other type is shallower and consists of smaller stones and gravel formed by similar agencies of physical weathering, but with a little finer material produced by chemical breakdown, part of which may be redistributed by deflationary action.

Gley podzols

Under the influence of an acid water table in the subsoil, gley podzols extend into the alpine environment as a physiographic climax.

In a typical profile the topsoil consists of a brown to brownish-black A_0 horizon on a grey to black mineral soil becoming lighter in colour with depth. The subsoil, frequently but not always of heavier texture, may be differentiated into a dark brown B_1 horizon of humus accumulation and a yellowish-brown to yellowish-red B_2 horizon of sesquioxides in the case of organo-gley podzols; or more usually consists only of the latter horizon of sesquioxide accumulation as in the drier sidero-gley podzols. The characteristic gley horizon of bluish-yellow mottle is usually present in the deeper subsoil.

Silty bog soils

The group of soils in the Australian Alps termed silty bog soils (eq. acid marsh soils) appears to be synonymous with the silty bog soils of Glinka, the meadow bog soils of Zakharov (Joffe, 1936), some of the half-bog soils of the U.S.D.A. (1938), and probably the semi-mineral limnic peats of Harris and Filmer (1947) in New Zealand. In so far as these soils are confined to alluvial or colluvial parent materials permanently influenced by acid ground- or surface-waters, they may be regarded as a physiographic climax.

The silty bog soils are characterized by a brown to brownish-black organo-mineral surface soil becoming paler and more mineral in composition with increasing depth. The organic and mineral fractions are usually so intermingled (due to the periodic addition of suspended mineral matter) that a distinct A_0 horizon is rarely well developed. If the water table is at or near the surface there is little or no horizon differentiation other than incipient gleying, but if the water table occurs at greater depths gley podzol horizons may also develop below the organo-mineral topsoil. The acid ground- and surface-waters and the high contents of silt and fine sand in these soils impart a characteristic silty texture and unstable crumb to single grain structure. This lack of good structure contrasts strongly with the pedogenically related but less acid normal meadow soils.

Snow patch meadow soils

The snow patch meadow soils are probably the most restricted group in Australia, not only being confined to the alpine tract but also requiring the presence of a semi-permanent snow patch on the immediate higher slope. Such snow patch environments are found on leeward aspects where snow accumulates to the greatest depth and persists for not less than about eight months of the year.

The snow patch meadow soils are the most acid and usually the shallowest subgroup of the meadow soils. In common with most members of this great soil group, they are characterized by a brownish-black to black organo-mineral horizon of good crumb structure becoming lighter coloured and more mineral in composition with increasing depth; depending on the average level of the water table, gleying usually develops in the surface soil or subsoil. They are also notably richer in clay and organic matter than adjoining alpine humus soils, and possess clays which are generally more kaolinitic. These differences in the amount and composition of the clay may be attributable to the mud and wind-blown dust received by the snow patch meadow soils from the melt-waters of the adjoining snow patch which acts as an absorbing and concentrating surface.

Bog peats

The bog peats of the alpine tract (including the valley bog and raised bog peats but not blanket bog peat) occur only in permanently wet, acid, and base-deficient situations: in the case of the raised bog peats the waters are soligenous whilst in the valley bog peats they may be either limnogenous or geogenous. That this physiographic moisture control is not absolute, however, is shown by the partial dependence of the peat-forming moss, *Sphagnum cymbifolium*,* on ombrogenous moisture; this is the dominant moss of the raised and valley bogs with which the corresponding peats are co-extensive. Details of bog structure and development in New South Wales are given by Costin (1954).

The characteristic profile features of the bog peats are due mainly to the remains of the acidophilous mosses from which they are partly or largely formed. The valley bog peats are intermediate between the fen and raised bog peats, consisting typically of rather compact, semi-decomposed peat of sedges and mosses, usually *Carex gaudichaudiana* and *Sphagnum cymbifolium*. In areas of locally higher pH, species of *Polytrichum* may replace *Sphagnum cymbifolium* as the dominant peat-forming moss. They may rest on a shallow layer of fen peat, below which basal silt or gravel, sometimes showing gley or podzol features, is encountered.

The raised bog peats are less compact, owing to the preponderance of mosses, usually *Sphagnum cymbifolium*. In the less decomposed deposits, evidence of the co-seral hollow-hummock bog succession is provided by the vertical sequence of plant remains which may still be preserved. Like the valley bog peats, the raised bog peats may be underlain by a thin layer of fen peat, and in the case of the shallower profiles, the basal mineral matter may be gleyed or podzolized.

Poor fen peats

In level or gently sloping situations under conditions of permanent limnogenous or geogenous wetness, poor fen peats are formed as a physiographic climax. Acidity as such does not appear to be critical

* The Australian representatives of *Sphagnum cymbifolium* (*sens. lat.*) are now referred to as *S. cristatum* Hpe.

for their genesis, since they may develop under quite acid conditions provided the associated fen vegetation receives an adequate base supply as suspended mineral matter.

In contrast to the largely decomposed rich fen peats, the poor fen peats are characterized by the extreme compactness and excellent state of preservation of the plant remains. The sedge *Carex gaudichaudiana* is by far the most important peat-forming species, with *Montia australasica*, *Plantago muelleri*, and *Astelia* spp. sometimes of local significance.

Humified peats

The physical environment of the humified peats resembles that of the bog peats and to a lesser extent the fen peats from which they have been derived, with the essential difference that conditions of relatively free drainage have been superimposed. Where drainage has been caused by natural agencies (e.g. climatic change, entrenchment of streams), these soils may be regarded as a climatic climax; where human interference has been responsible they represent an anthropogenic climax.

The humified peats of the alpine tract resemble the parent peats in being dominated by organic or organo-mineral horizons resting abruptly on basal mineral matter, and alpine humus soils in their excellent crumb structure due to the abundance of humus. Gleying and podzolization may occur in the mineral layers and sometimes also at the base of the lowest organic horizon.

Interrelationships of Alpine Soils in Australia

The interrelationships of alpine humus soils in New South Wales and those of lower levels are considered in other publications (Costin, Hallsworth, and Woof, 1952; Costin, 1954). In this paper mention is made only of the topographic interrelationships of the alpine soils themselves. These are best illustrated by considering two general sequences from top to bottom of the slope, upon which micro-catenas may be superimposed depending on minor variations in topography.

The major catenas which develop under acid ground-water conditions on gneissic granite in the alpine tract at Kosciusko vary somewhat in accordance with the position of the water table. Where the water table on the flat is at or near surface level for most of the year, resulting in wet, strongly acid, and base-deficient conditions, the sequence from top to bottom of the slope is alpine humus soil (tall alpine herbfield and sod tussock grassland) in all freely drained upper, middle, and lower catenary positions, raised bog peat (raised bog) on the wet lowest slopes, and valley bog peat (valley bog) and poor fen peat (fen) on wet situations on the flat under strongly acid and acid conditions respectively. Where the level of the water table near the flat is not so high, the raised bog peats and valley bog peats respectively are replaced by gley podzols and silty bog soils, with a vegetation of sod tussock grassland. (Fig. 1.)

In very exposed upper catenary positions the alpine humus soils of the previous macro-catenas may be replaced by lithosols with fjaeldmark vegetation. Where this upper position is a leeward aspect accumulating snow, lithosols with other fjaeldmark communities occur above the snow

patch and grade into snow patch meadow soils with short alpine herbfield at the base.

The most noteworthy topographic sequence involving snow patch meadow soils is the micro-catena from the wettest, usually flat and most central area below the snow patch to the drier, adjacent, relatively snow-free sites. This sequence is shallow poor fen peat (short alpine herbfield), snow patch meadow soil (short alpine herbfield), and alpine humus soil (tall alpine herbfield).

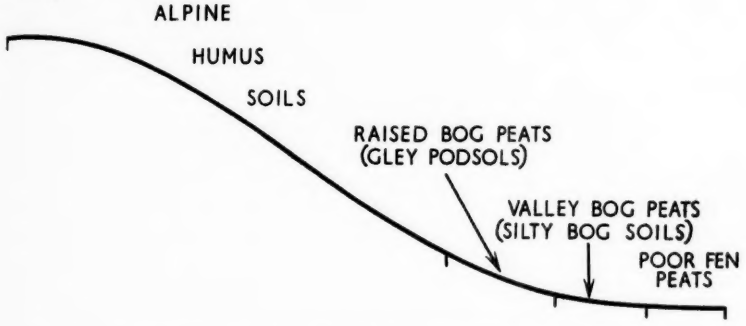


FIG. 1. Topographic sequence, Kosciusko. (Brackets indicate variations in sequence due to lower water table near the flat.)

A common micro-sequence from the centre to the edge of raised bogs is raised bog peat, raised bog gley, raised bog gley podzol, and gley podzol.

Another interesting micro-sequence, which may occur within a distance of a few yards as the influence of the surface waters becomes progressively less and the level of the water table deeper, is silty bog soil, silty bog gley podzol, and gley podzol, commonly with a vegetation of sod tussock grassland throughout.

The spatial relationships of the peats, silty bog soils, and gley podzols may be interpreted in terms of the pH and base and mineral content of the ground-waters. With increased base content, with or without change in acidity, the raised and valley bog peats are replaced by poor fen peats, or, if sufficient suspended mineral matter is being added, by silty bog soils. If the level of the water table falls, gley podzols, not silty bog soils, are formed.

In recently drained areas all intermediate stages from humified peat to bog peat may be found; where a longer period of drainage has led to the complete humification of the original peat, the humified peat usually adjoins alpine humus soils.

The Alpine Humus Soils of Australia in Relation to Homologous Groups Overseas

In the alpine tract of the Australian mainland is found a combination of historical, physiographic, climatic, and biological conditions apparently not encountered overseas. The relative mildness of the Pleistocene

glaciations and the length of the post-glacial period, the comparatively gentle topography, the contrasting winter and summer climates, and the influence of the grass *Poa caespitosa* and the mountain earthworm *Megascolex* sp. have favoured the development of alpine humus soils to an unusual degree.

Not only are most of the New Zealand and European mountains steeper than the Australian Alps by virtue of their origin as fold mountains, but they have also been affected by glaciations of greater duration and severity. The main pedogenic implications of this are that these alpine soil parent materials were predominantly steep and rocky and have been available for the formation of present-day soils for considerably shorter periods.

In many high mountains where precipitations increase and temperatures decline with increasing altitude there is apparently a narrow tract where the moisture-temperature combination is optimum for soil and plant development (cf. Costin, Hallsworth, and Woof, 1952). Under such conditions, low temperature limits development above this optimum tract, irrespective of increased moisture, whereas inadequate precipitation limits it below. Where the mountains are very steep, therefore, the area of the optimum moisture-temperature combination will be only small. Where the mountains are gentler, however, the area of optimum development will be correspondingly greater, and in the case of plateau mountains may be very great indeed. The latter condition appears to obtain over much of the Australian Alps.

While biological conditions may so dominate the alpine soil environment in some countries as to produce peats or podzolized soils, they do not appear to operate in stabilizing soil sesquioxides and bases to the same extent as the snow grass and mountain earthworm of the Australian Alps.

These points may now be illustrated by a brief outline of some homologous alpine soils in other countries.

In the alps of New Zealand, alpine humus pedogenesis appears to be limited by physiography and climate. Precipitous slopes, especially in the dry eastern alps of the South Island, favour lithosol formation, and alpine humus soils, where developed, are rarely deep. On the western alps, furthermore, excessive precipitations tend towards the formation of ombrogenous soils resembling blanket bog peats. In the volcanic alps of the North Island, where the climate appears more favourable for alpine humus pedogenesis, the last eruptions have been so recent that little profile differentiation has occurred; bare rock and lithosols thus dominate the landscape.

The limited alpine environments of Britain are distinguished by their moist, maritime climate often combined with strong winds. For the most part this has favoured the genesis of blanket bog peats and to a less extent podzolized soils, with lithosols on wind-exposed summits. Under special conditions of parent material and slope, however, unpodzolized organo-mineral soils are encountered, as on the mica schists of Ben Lawers in Scotland and the gabbros of the Black Cuillins on Skye. Given this suitable combination of basic parent material and active soil

creep, the peat-forming and podzolizing processes which dominate the British mountains generally are locally suppressed. The resultant soils, known as creep soils and similar to the so-called truncated podzols of Robinson (1949), may be regarded as an acid colluvial brown earth. They are the closest approach to the alpine humus soils in Britain.

The Scandinavian alpine areas, with their very wide range of environments, present a corresponding array of alpine soils, but in no area with which the writer is familiar, except perhaps for the Jotenheimens in Norway, are well-developed alpine humus soils encountered. Under very oceanic conditions near the Norwegian fiords, for instance, blanket bog peat, lithosols, and bare rock are characteristic. By contrast, the less precipitous drier alpine areas of Swedish Lappland, where the soil is frozen in winter, are more favourable for the genesis of brown earth-like soils and rendzinas on relatively basic rocks, with podzolized soils on more acid parent materials. The persistence of the former unpodzolized types depends on the effects of high ground-waters due to an immature drainage system, combined with rapid natural erosion and intense solifluction. Owing to the shortness of the arctic-alpine growing season, insufficient organic matter is produced in these soils to give them an alpine humus character.

In the type areas in the Continental Alps in which they were first studied in detail (Braun-Blanquet and Jenny, 1926), the alpine humus soils are typically shallow and restricted to more acid parent materials at alpine levels. In these areas soils resembling brown limestone soils and rendzinas occur on basic parent materials, whilst the alpine humus soils on acid rocks give way to podzolized types at lower levels. Under moister alpine conditions the relation between soil and parent material still obtains, but less so, with the soil formation on basic rocks tending towards alpine humus pedogenesis, and the altitudinal range of the alpine humus soils on silicate parent materials slightly extending. Finally, under the moistest alpine conditions, the genesis of alpine humus soils may in favourable situations extend also to basic rocks. In general, however, these soils are shallower than those in Australia and subject to geological control.

Proceeding from the Continental Alps through the Maritime Alps across to the East and Central Pyrenees, the topography, while still often precipitous, tends to become less so; there is a strong contrast between the winter and summer seasons; added to which the glacial history has probably been less severe. These features express themselves in parts of the Pyrenees in an alpine environment which has a general similarity to the alpine areas in Australia, including a strong development (where not eroded) of alpine humus soils. Even in the Pyrenees, however, the alpine humus soils are not as well developed, either as regards profile depth or altitudinal range, as those in Australia: a difference which may be attributed to a less favourable moisture-temperature combination for pedogenesis combined with a less vigorous biological environment.

It would thus appear that the combination of soil-forming factors tending towards alpine humus pedogenesis has been most favourable in the Australian Alps, where it has resulted in the extension of this great

soil group as the climatic climax over a wider range of altitude, slope, and parent material than appears to be the case in other countries.

The Utilization of Australian Alpine Soils

Most of the alpine country of the Australian mainland has been used for summer grazing of sheep and cattle for periods approaching 100 years, at first indiscriminately and during more recent years under snow-lease systems. Cattle have been replaced by sheep as the dominant grazing animal in the Snowy Mountains, but in the Victorian Alps the former livestock still predominates. Associated grazing practices have included burning and to a less extent roadmaking and draining. Most recently in the Snowy Mountains, engineering practices associated with water conservation schemes threaten to become the major disturbing influence.

From the information given previously on the alpine environment in Australia it might at first be anticipated that the Australian Alps, more so than most others, would be resistant to anthropogenic disturbance; indeed, the favourable glacial and post-glacial history, the climate, the gentle topography, the deep soils, and the unusual biological conditions might be taken as *a priori* evidence. On closer examination, however, it is found that the Australian alpine environment is most susceptible to disturbance, and for the very reasons listed above.

In the first place, the very favourable conditions for pedogenesis have produced what are essentially soil mountains, in strong contrast to the rock and peat mountains which characterize most other alpine environments. In mountains of the former type, the erosion risk is clearly greater than in the latter. The comparatively long and unbroken slopes of the Australian Alps not only increase the erosion risk but render ground-water soils and vegetation on lower slopes more dependent on the stability of the slope as a whole.

The wide daily temperature fluctuations of the snow-free months appear to constitute a far greater erosion hazard than in most homologous low-alpine areas. In the latter areas, severe daily freezing and thawing are not as common; nor are its effects as serious on rock or peat as on surfaces of soil. The efficiency of Australian alpine catchments and especially the Snowy Catchment, furthermore, partly depends on the presence of bogs and other ground-water soils and vegetation (Costin, 1952). In view of the warm summers, however, the present alpine environment on the Australian mainland appears barely suitable for the active development of the bog moss *Sphagnum cymbifolium*, so that, once damaged, the bogs are in danger of progressive destruction.

Finally, the more severe Pleistocene history of Europe and the evolutionary influence of close-grazing herbivores have favoured a stronger development of protected renewal buds in its flora. The alpine vegetation of Europe, therefore, appears to be better adapted to grazing by domestic livestock than the plant communities of the Australian Alps.

Bearing these points in mind, it is not surprising that considerable deterioration of the soils and vegetation has already occurred in alpine areas of the Australian mainland. The nature and extent of this damage

lie beyond the scope of this paper; they have been reported elsewhere (Costin, 1950, 1952, 1954).

Summary

The alpine soils of the Australian mainland are considered in relation to alpine soils in homologous regions of Europe and New Zealand.

The environmental factors of the Australian Alps combine in a unique manner to produce what may be termed soil mountains compared with the predominantly rock or peat mountains of most other alpine regions.

The Australian alpine soils include eight great soil groups (with six subgroups): alpine humus soils, lithosols, gley podzols (sidero-gley podzol and organo-gley podzol), silty bog soils, meadow soils (snow patch meadow soil), bog peats (valley bog peat and raised bog peat), fen peats (poor fen peat), and humified peats. These soils are briefly described and their interrelationships are indicated.

The factors which combine so favourably for alpine humus pedogenesis in Australia are examined further in relation to their pedogenic effects in Europe and New Zealand.

This combination of factors is also found to render Australian alpine areas unusually vulnerable to anthropogenic disturbance, as by grazing and engineering operations.

Acknowledgement

Grateful acknowledgement is made to the Trustees of the Australian Services Canteens Trust Fund for the scholarship which enabled part of these investigations to be undertaken.

REFERENCES

Beadle, N. C. W., and Costin, A. B. 1952. Ecological classification and nomenclature. Proc. Linn. Soc. N.S.W. **77,** 61.

Braun-Blanquet, J., and Jenny, H. 1926. Vegetationsentwicklung und Bodenbildung in der alpinen Stufe der Zentralalpen. Neue Denkschr. Schweiz. Naturf. Ges. **63,** 175.

Costin, A. B. 1950. Mass movements of the soil surface with special reference to the Monaro region of N.S.W. J. Soil Cons. Ser. N.S.W. **6,** 41, 73, 123.

—— 1952. Hydrological studies in the upper snowy catchment area with special reference to the effects of land utilization. J. Soil Cons. Ser. N.S.W. **8,** 5.

—— 1954. A Study of the Ecosystems of the Monaro Region of New South Wales. Govt. Printer, Sydney.

—— Hallsworth, E. G., and Woof, M. 1952. Studies in pedogenesis in N.S.W. III. The alpine humus soils. J. Soil Sci. **3,** 190.

Hallsworth, E. G., and Costin, A. B. 1950. Soil classification. J. Aust. Inst. Agric. Sci. **16,** 84.

Harris, W. F., and Filmer, D. W. 1947. A peat profile in the agricultural area, Hauraki Plains, New Zealand. D.S.I.R. (N.Z.) Botany Div. Publ. No. 12.

Joffe, J. S. 1936. Pedology. Rutgers, New Jersey.

Robinson, G. W. 1949. Soils, Their Origin, Constitution and Classification. Murby, London.

U.S.D.A. 1938. Soils and Men. U.S. Govt. Printer.

APPENDIX

Profile Characteristics of Some Alpine Soils

Alpine Humus Soils

Profile No. C21
Profile No. S18 } Costin, Hallsworth, and Woof (1952).

Gley Podzols

Profile No. CS2 (*Organo-gley Podzol*)

Environment. Alpine tract, near Snowy River Bridge, Kosciusko; elevation 6,270 ft.; level; imperfect drainage; parent material gneissic granite alluvium; sod tussock grassland vegetation of *Poa caespitosa–Danthonia nudiflora* alliance; average annual precipitation about 85 in., including heavy winter snow.

A_0 0–3 in. Light brown, semi-decomposed, slightly laminated leaf litter of *Poa caespitosa* (ecotype) with a little moss merging with A_1; organic carbon 8·1 per cent.; pH 4·7.

A_1 3–7 in. Brownish fine silty loam, merging with A_2; fine mineral skeleton of biotite; single grain structure; fine porous to closed, loose; a little humus and few roots; moist; occasional earthworms; clay 15·4 per cent.; organic carbon 4·5 per cent.; pH 4·8.

A_2 7–12 in. Pale yellowish-grey loam, fairly well defined from B_1; mineral skeleton of biotite and a little quartz gravel; single grain structure; fine porous to closed, loose; very little humus, a few roots; moist; occasional earthworms; clay 10·2 per cent.; organic carbon 1·8 per cent.; pH 4·9.

B_1 12–20 in. Greyish-black silty clay loam, merging slightly with B_2GC; fine mineral skeleton of biotite; loose crumb to single grain structure; friable to compact; translocated humus moderate in amount, very few roots; damp; no fauna observed; clay 21·9 per cent.; organic carbon 3·4 per cent.; pH 4·7.

B_2GC 20–30 in. Yellowish gravelly clayey loam becoming increasingly gravelly and stony below 30 in.; mineral skeleton of biotite with moderate amount of quartz and granite gravel and stones; loose crumb to single grain structure; slightly porous due to coarse mineral skeleton, compact to slightly tenacious; no visible humus or roots; damp to wet; no fauna observed; clay 5·3 per cent.; organic carbon 0·8 per cent.; pH 5·1.

Silty Bog Soils

Profile No. RK1 (*Silty Bog Gley Podzol*)

Environment. Alpine tract, Kosciusko; elevation 6,500 ft.; level; impeded drainage; mixed organic and colluvial inorganic parent material in drainage area of gneissic granite; fen vegetation of *Carex gaudichaudiana* alliance; average annual precipitation about 85 in., including heavy winter snow.

H–A_1 0–5 in. Brownish-black, organo-mineral silty clay loam, merging with more mineral A_2; abundant fine mineral skeleton of biotite; almost single grain structure; fine porous to closed, slightly tenacious; humus and roots abundant; moist; occasional earthworms; sand 36·7 per cent.; silt 17·7 per cent.; clay 29·1 per cent.; organic carbon 9·6 per cent.; pH 4·7.

A_2 5–19 in. Ashy-grey silty clay loam, fairly clearly defined from BGC; abundant fine mineral skeleton of biotite; single grain structure, except for slight stratification due to original conditions of deposition; closed, tenacious; very little humus, few roots; moist to damp; no fauna observed; sand 33·4 per cent.; silt 26·8 per cent.; clay 28·1 per cent.; organic carbon 6·8 per cent.; pH 4·8.

BGC 19–30 in. Bluish-yellow mottled gravelly sandy loam, with brown staining at 20 in. due to translocated humus; abundant mineral skeleton of biotite and rotten gneissic granite; single grain structure; almost closed, tenacious to semi-indurated due to abundance of stones and gravel; very little humus and very few roots; damp to wet, with water table at 30 in.; no fauna observed; sand 73·9 per cent.; silt 16·0 per cent.; clay 9·1 per cent.; organic carbon 0·6 per cent.; pH 4·9.

Snow Patch Meadow Soils

Profile No. 1X

Environment. Alpine tract, Mt. Kosciusko; elevation 7,100 ft.; moderate slope, easterly aspect, situated below semi-permanent snow patch; slightly impeded drainage; colluvial parent material of gneissic granite; short alpine herbfield of *Plantago muelleri–Montia australasica* alliance; average annual precipitation about 90 in., including heavy winter snow.

A_0–A 0–6 in. Brownish-black sandy clay, peaty near surface but becoming increasingly mineral towards mergence with GC; a small quantity of quartz and gneissic granite gravel and stones; crumb structure; fine porous, slightly friable to compact; humus and roots very abundant; damp to wet; occasional earthworms; organic carbon 12·8 per cent.; pH 4·6.

GC 6–24 in. Brownish-grey to bluish-grey gravelly clay loam, becoming coarser and showing slight gleying towards underlying rock at 24 in.; single grain structure; fine porous, harsh feel due to gravel and stones; little humus and few roots; wet; no visible fauna.

Bog Peats

Profile No. d (*Raised Bog Peat*)

Environment. Alpine tract, Mt. Kosciusko; elevation 7,050 ft.; gentle slope, north-easterly aspect; permanently wet; organic parent material on gneissic granite; raised bog vegetation of *Epacris paludosa–Sphagnum cymbifolium* alliance; average annual precipitation about 90 in., including heavy winter snow.

0–3 in. Yellowish-green, undecomposed *Sphagnum* moss; porous, spongy, wet; organic carbon 43·7 per cent.; pH 4·0.

3–6 in. Brownish to black, partly decomposed moss peat; layered structure; fairly open, greasy feel; saturated with water; organic carbon 45·6 per cent.; pH 3·8.

9–15 in. Brownish-black to black, largely decomposed moss peat; almost structureless, with original layered structure barely preserved; fairly open, greasy feel; saturated with water; organic carbon 38·7 per cent.; pH 3·8.

15–18 in. Black, thoroughly humified peat; single grain structure; slightly closed, greasy feel; saturated with water; organic carbon 38·0 per cent.; pH 4·3.

18–22 in. Coarse gravel resting on gneissic granite at 22 in.

Poor Fen Peats

Profile No. T1

Environment. Near Mt. Townsend in the alpine tract, north of Mt. Kosciusko; elevation approx. 7,000 ft.; level; permanently wet; organic parent material; fen vegetation of *Carex gaudichaudiana* alliance; average annual precipitation about 90 in., with abundant winter snow.

0–12 in. Brownish *Carex* peat, loose near surface but becoming increasingly compact; plant remains still visible, mixed with a little mineral matter; layered structure; roots abundant; wet but not water saturated; organic carbon 22·3 per cent.; pH 4·6.

12–24 in. As above, but more compact and with thin bands of mineral matter at 20–24 in.; organic carbon 19·2 per cent.; pH 5·4.

24–36 in. Black silty gravel; greasy feel; becoming slightly coarser with increasing depth; organic carbon 3·8 per cent.; pH 5·4.

Humified Peats

Profile No. BG1

Environment. Kosciusko, at headwaters of Guthrie's Creek; elevation 5,900 ft.; upper subalpine near transition to alpine tract; gentle slope; satisfactory drainage in organic horizons but water table present in underlying mineral matter; parent material raised bog peat on gneissic granite; seral vegetation of *Calostrophus lateriflorus* being replaced by subalpine woodland; average annual precipitation about 70 in., with abundant winter snow.

H–A_1 0–13 in. Black silty clay, merging with H–A_2; greasy feel; mineral skeleton almost lacking; original peat largely humified; fine crumb structure; fine porous, mellow; humus and roots abundant; moist; mountain earthworms common; clay 35·8 per cent.; organic carbon 13·2 per cent.; pH 4·6.

H–A_2 13–22 in. Greyish-black silty clay loam, with slight bleaching at 20–22 in., clearly defined from B; greasy feel; mineral skeleton almost lacking; original peat largely humified; fine crumb to single grain structure; almost closed, slightly tenacious; humus and roots abundant; damp; mountain earthworms present; clay 24·0 per cent.; organic carbon 6·9 per cent.; pH 5·1.

B 22–33 in. Yellow silty loam; stones and gravel abundant; single grain structure; almost closed, sticky and tenacious; few roots and no humus; wet, with water table at 33 in.; clay 15·8 per cent.; organic carbon 1·0 per cent.; pH 5·3.

SOME SOIL-FORMING PROCESSES IN THE HUMID TROPICS

II. THE DEVELOPMENT OF THE UPPER-SLOPE MEMBER OF THE CATENA

P. H. NYE
(*University College of the Gold Coast*)

Introduction

In an earlier paper (Nye, 1954) a description was given of a catena developed over a biotite granitic gneiss in the West African forest region at Ibadan, south-west Nigeria. Since the lower-slope members of the catena are formed largely from material which has crept down from the upper slopes, a complete understanding of the development of the catena must begin with a study of the upper-slope member. In this paper, therefore, the results of a detailed study of a mature upper-slope profile nearly 30 ft. deep will be presented. All stages in the formation of the profile from fresh rock have been considered, including the development of the parent material of the solum.

Description of the Profile

Vegetation: Secondary woodland, about 20 years old.

$+1\frac{1}{2}$–0 in.	Dark brown (7·5 YR 4/2) loose worm casts and broad leaf litter.
0–$\frac{1}{2}$ in. Cr W	Humic dark brown (7·5 YR 4/2) fine sandy clay loam from casts; weak crumb; sharp change to
$\frac{1}{2}$–6 in. Cr T	Dark reddish-brown (5 YR 3/3) humic loose loamy coarse sand; single grain; gradual change to
6–15 in. Cr G_1	Reddish-brown (5 YR 4/3) friable gravelly loamy coarse sand; gravel of angular quartz and hard purple to softer yellowish-brown, iron pisolites; diffuse change to
15–26 in. Cr G_2	Red (2·5 YR 4/6) gravelly sandy clay loam; compact; structureless; few roots; gradual change to
26–72 in. S_1	Red (2·5 YR 4/8) coarse sandy clay; compact; structureless; few small red and yellow pockets of decomposing gneiss grading into pisolites above 48 in.; diffuse change to *oxidation-reduction zone*:
72–120 in. S_2	Decomposed gneiss of reddish-yellow powdery porous kaolinized felspar bands and red decomposed biotite-rich bands; abundant light grey, mottled red, compact sandy clay channels and pockets, often with traces of roots at their centre; diffuse change to
10–20 ft. S_{31}	Decomposing gneiss of white powdery porous kaolinized felspar and quartz, and grey, decomposing to strong brown and red, biotite-rich bands; friable and very easily dug; some compact light grey sandy channels above 16 ft. and fine holes to 23 ft.; diffuse change to the *reduced zone*:
20–28 ft. S_{32}	Porous decomposed gneiss showing only white and grey bands. Water seepage at 28 ft. prevented further digging.

Laboratory Examination

The following determinations were carried out on samples collected from all creep horizons, and from the sedentary horizons at intervals of 6 in. from 2–4 ft., 1 ft. from 4–6 ft., 2 ft. from 6–16 ft., and 4 ft. below 16 ft.

Mechanical analysis (Table 1). By the pipette method. Preliminary crushing was avoided, and by washing on a 0·5-mm. sieve under a tap all soft incipient concretions were satisfactorily preserved while coarse aggregates were destroyed. The coarse earth was divided into fractions > 7, 7–4, 4–2, 2–1, 1–0·5 mm., and the relative proportions of mineral grains and concretionary sand or gravel in each fraction were determined roughly by counting 50 individuals.

Exchange properties (Table 1). Exchangeable cations were determined by the micro-methods of Peech (1945); exchangeable acidity by Schofield's (1933) para-nitrophenol method; and pH with the glass electrode using a 2·5 : 1 water to soil ratio.

Free iron oxide in the clay (Table 1). The clay was separated by decantation, and iron oxide extracted with hydrosulphite.

TABLE 1

Selection from the Analytical Data

	Cr W	*Cr T*	*Cr G*$_1$	S_1	S_2	S_{31}	S_{32}
Horizon depth (in.)	0–1	2–3	6–12	30–36	72–96	12–15 ft.	23–24 ft.
Size fraction (mm.):							
> 4% (concretions in brackets)	1	2(1)	36(28)	7(2)	2(1)	..	2
4–0·5%	25(3)	49(7)	49(10)	40(14)	26(8)	15	36
0·5–0·2%	32	26	22	9	9	11	16
0·2–0·02%	23	13	14	9	18	25	30
0·02–0·002%	7	5	6	6	13	18	13
< 0·002%	13	7	9	36	34	31	5
pH	6·2	6·4	6·7	6·6	5·5	5·5	8·0
Ex. cations (m.e./100 g.):							
Ca	n.d.	2·5	1·9	2·2	1·7	1·4	2·5
Mg	..	1·3	0·8	1·1	0·9	1·1	1·9
K	..	0·43	0·34	0·32	0·34	0·34	0·36
Na	..	0·49	0·50	0·49	0·30	0·43	0·46
Acidity	..	2·0	1·6	2·0	2·9	2·9	0·07
Org. C. %	0·96	0·53	0·32	0·31	..	..	..
Total Fe_2O_3%	3·3	3·0	3·7	6·5	4·8	5·2	2·3
Free Fe_2O_3 in clay %	9·4	8·7	8·4	8·2	7·2	5·2	1·3

Organic carbon (Table 1) was determined by the Walkley and Black (1947) method.

Mineralogical studies (Table 3). The fine sand (0·2–0·02 mm.) and medium sand (0·5–0·2 mm.) were pretreated with sodium hydrosulphite and mounted by the method of Marshall and Jeffries (1946). Counts of 200 grains were made on each fraction.

Identification of clay minerals by X-ray diffraction on glycerol-treated orientated aggregates, using a 9-cm. camera and CuKα-radiation.

The General Evolution of the Profile

(*a*) *The alteration of the rock.* Merrill (1897), Harrison (1934), Goldich (1938), and Brock (1943) have analysed exposures of fresh rock, and the increasingly weathered rock which overlies it or lies near by. It seems likely in these instances that the weathering has largely occurred under oxidizing conditions. There appears to have been no study made of the alteration of rocks in the zone of saturation; yet, as Campbell (1917) has pointed out, decomposition, particularly of the ferromagnesian minerals, will be very different in these two conditions. He distinguishes between 'alteration' taking place in the zone of permanent saturation and 'weathering' taking place above that level.

In the present study the rock exposed above the water-table was highly decomposed even though it showed all the structure of the fresh rock. It was porous, very easily dug, and, when dry, powdery. Its volume weight, determined by weighing a dry clod, coated thinly with wax in air and water, was 1·53 g. per ml. Thus at least 41 per cent. of the original material in the fresh rock has been lost by solution in the ground water. Further decomposition has provided 6 per cent. of clay. It may be shown that this substantial loss of material has been borne mainly by the felspars, the biotite being changed to vermiculite only. In crushed fresh rock the percentages of quartz, felspar, and biotite were about 28, 62, and 10 respectively, while in the decomposed rock above the water-table they are roughly 43, 14, and 43, the last figure being high because the biotite grains tend to flake in separation. However, from the ratio of quartz to felspar it may be calculated that 85 per cent. of the original felspar grains have decomposed.

The alteration of the rock probably begins at considerable depth, on the surfaces of narrow water-bearing fissures. As decomposition proceeds the water will penetrate into the rock, until at the surface of the water-table each mineral grain is in contact with a film of water. Most of the decomposition products pass into the drainage system, with the exception of the small amount of clay formed. The experience of water-supply geologists (Tattam, 1943; Meinzer, 1923) suggests that the water at 28 ft. represents a perched water body held up near the surface of the fairly fresh rock, whose joints are plugged with clay. The regional water-table is probably formed in the joints of the fresh rock at a considerable depth.

There is little change in the altered gneiss or its minerals from the water level at 28 ft. to a depth of 20 ft. The lower part of this reduced zone will be the capillary fringe.

(*b*) *The zone of weathering.* The change from the zone of reduction to the zone of oxidation-reduction* at about 20 ft. is shown by oxidation of

* The expression 'zone of oxidation-reduction' is preferred to 'zone of intermittent saturation' used by Campbell, since the latter term could suggest a fluctuating water-table, while all that it is intended to suggest here is that there are reduced and oxidized patches in the soil.

the biotite bands in the gneiss. There is, however, little diffusion of iron into the bands of kaolinized felspar and quartz, which are white below a level of 10 ft.

An interesting and important phenomenon is associated with the roots which occur sparsely in the lower part of this zone. Occasional channels of compact pale grey sandy clay up to $\frac{1}{2}$ in. in diameter may be seen cutting across the planes of foliation of the gneiss. They usually have traces of roots at their centre. The pale grey colour and smooth compact appearance of these channels is quite different from the white porous powdery appearance of the kaolinized felspar and quartz in the bulk of the soil. Vine (1949) has noted these pockets in other parts of Nigeria. A comparison of the mechanical and mineralogical composition of these channels with those of a reduced and oxidized patch in the surrounding soil at 15 ft. shows that the only essential difference lies in the very great increase in the clay content of the channel (Table 3). The relative proportions of the minerals do not indicate any greatly accelerated decomposition around the root, so that the extra clay appears to be almost entirely illuvial, having been washed down the root channel into the porous soil.

From 10 ft. to 6 ft. there is a change from an horizon of decomposed gneiss with numerous pockets of illuvial sandy clay into an illuvial horizon of compact sandy clay containing numerous residual pockets of decomposed gneiss. The volume weight has increased to 1·60. The sandy clay is now coloured grey and red. The biotite bands are red and the felspar bands are yellowish-red.

Above 6 ft. there are no reduced colours, and the soil is a compact red concolorous coarse sandy clay. The volume weight of a clod at 5 ft. has risen to 1·81. A few small residual pockets of decomposed gneiss remain, which tend to harden to concretions towards the top of the sedentary horizon at 26 in., a process discussed fully below.

Above 26 in. there is the horizon of soil creep, whose development and subdivision into gravel, termite, and worm horizons has already been considered in Part I

Against this general background the different aspects of the evolution of the profile may now be discussed in some detail.

A. *The Weathering of the Constituent Minerals of the Gneiss*

The gneiss consisted of light-coloured bands of coarse-grained quartz and felspar up to about 2 in. wide, alternating with narrower, finer-textured, highly schistoze, dark-coloured bands of quartz, felspar, and biotite. The felspar was mainly microcline, with subordinate plagioclase of approximate oligoclase-andesine composition.

A typical selection from the mineralogical composition of the 0·5–0·2 mm. and 0·2–0·02 mm. sand fractions of the profile samples is shown in Table 2. Particles larger than 0·5 mm. were nearly all quartz.

(*a*) *Weathering of quartz.* At 28 ft. and 23 ft. much of the quartz is very cracked and fissured and is sharply angular. At 15 ft. fissures are not nearly so pronounced, suggesting that the larger fragments have broken into more stable unfissured grains. At 8 ft. the grains are subangular

TABLE 2

Mineralogical Composition of the Sand Fractions (per cent.)

Depth	*Quartz*	*Felspar*	*Biotite*	*Iron oxide*	*Zircon*
		(*a*) *Medium Sand* (0·5–0·2 *mm.*)			
12–18 in.	94	1	..	4	..
72–96 in.	67	14	..	19	..
23–24 ft.	50	33	17	..	..
		(*b*) *Fine Sand* (0·2–0·02 *mm.*)			
12–18 in.	53	20	23	3	1
72–96 in.	15	18	66	..	1
23–24 ft.	16	7	77	..	..

TABLE 3

Mechanical and Mineralogical Analysis and pH of Clay Channels at 15 ft. compared with Surrounding Soil

	Coarse sand	*Fine sand*	*Silt*	*Clay*	*pH*
Clay channel .	13	14	11	62	5·0
Oxidized patch	29	34	18	19	5·1
Reduced patch	20	32	21	27	4·9

	0·5–0·2 *mm. fraction*			0·2–0·02 *mm. fraction*		
	Quartz	*Felspar*	*Biotite*	*Quartz*	*Felspar*	*Biotite*
Clay channel .	71	10	19	8	4	88
Oxidized patch	62	18	20	9	10	81
Reduced patch	60	8	32	4	5	91

and this form persists to the surface. The same rounding of the sharp edges is noticeable in the coarse sand and gravel. The surfaces of the grains when cleaned appear quite fresh and show no evidence of dissolution. Above 0·2 mm. quartz dominates all fractions, but there is a fairly sharp lower limit to the size of quartz grains in the rock, and in the fine sand fraction (0·2–0·02 mm.) the proportion of quartz is quite small; even in the top soil, where it is concentrated by the weathering of the other minerals, it forms only 56 per cent. of the fraction.

(*b*) *Weathering of felspar.* Much of the felspar in apparently fresh specimens of rock is seen under the microscope to be kaolinized. At 28 ft. the cleaned grains were somewhat kaolinized and their surface showed a dark brown pitting. At 6 ft. they were all kaolinized and much cracked along cleavages. The very considerable loss of felspar below the water-table has already been mentioned. Between the water-table and a depth of 6 ft. the ratio of felspar to quartz in both sand fractions remained fairly steady at about 1 : 3. Above 6 ft. the proportion in the medium sand fraction dropped sharply, possibly owing to breaking along cleavages, but in the fine sand fraction there was no consistent decline. There was considerably more potash felspar than plagioclase in all samples except the one at 28 ft., in which the proportions were equal.

These observations suggest that felspar which has emerged from the zone of alteration is relatively stable in the zone of weathering.

(*c*) *Weathering of biotite.* The decomposition of biotite in soils under oxidizing conditions has recently been studied by Walker (1949), who found that it altered directly to vermiculite; while Stephen (1952) found that it altered to vermiculite through an intermediate chlorite stage. In the present study the alteration in the reduced zone appeared to follow the course observed by Stephen, but in the oxidized zone was carried a stage farther.

The biotite in thin sections of the fresh rock was strongly pleochroic, and varied in colour from brown to olive; and in a sample of decomposed rock obtained from below the water-table in the valley the colour of all the flakes was olive. At the water level in the deep profile, alteration to golden-yellow vermiculite was well advanced. The flakes exfoliated readily when treated with cold 20 vol. hydrogen peroxide. All the surfaces were smooth and fresh-looking. At 15 ft., in the oxidation-reduction zone, some flakes showed a fine mottled surface with dark pitting, and some were dark brown or almost black, especially towards the centre. At the base of the oxidized zone at 6 ft. nearly all flakes were pitted, darkened, or black, and they became increasingly black towards the top of the profile. The blackening is probably due to a deposit of ferric oxide within the mica layers, for it is not removed by the normal hydrosulphite cleaning process. The black flakes are magnetic and sink in bromoform. The stability of the iron oxide liberated by the biotite is no doubt associated with the general stability of concretionary iron oxide in humid regions with a hot and dry season.

There was no biotite in the sand fraction above 0·5 mm. In the medium sand fraction altered biotite was well represented, though at 8 ft. all flakes were blackened, and in the upper layers the black flakes disappeared. In the fine sand fraction altered biotite was dominant in all samples below 3 ft. The proportion of fine sand in the mechanical analysis does, however, drop steadily from 32 per cent. at 20 ft. to 9 per cent. at 3 ft., and this loss must be largely due to weathering of the biotite. The loss in the silt fraction which is also largely altered biotite is equally marked. Above 3 ft. the proportion of altered biotite fell sharply.

B. *Analyses of the Clay Minerals* ($< 2\ \mu$.)

Orientated aggregate photographs show that at all levels the dominant mineral was a member of the kaolin group, and with it was associated about 5 per cent. of a hydromica. This was of the trioctahedral type, since in all photographs the 002 reflection was absent (Brown, 1951). There was no sign of a 14kX reflection corresponding to chlorite or vermiculite, which suggests that the mica is authigenic in origin. In contrast to the trioctahedral mica described by Walker (1950), it appears to be relatively stable.

At the water-table, where the pH is 6·8, there was about 25 per cent. of montmorillonoid in the clay fraction, but at 24 ft. there was only a trace, though the pH was still high at 8·0. It may be suggested that the

montmorillonoid requires for stability in this profile not only alkali and alkaline earth cations and high pH, but also sufficient silicate ion.

The relative amounts of the minerals were estimated visually from the intensity of the lines, using MacEwan's data (1945), the amount of mica being reduced by a factor of 3·5 to allow for the change from a di- to a tri-octahedral type.

A powder photograph of a sample from the top of the sedentary horizon showed kaolin, goethite, and quartz. There was no trace of other forms of iron oxide or of aluminium oxide.

C. *The Development and Properties of the Iron Oxide Concretions*

Casual inspection suggests that the concretionary gravel is being formed or has been deposited in the gravel accumulation horizon; but close examination shows that it actually originates in the upper layers of the sedentary horizon by deposition and hardening of iron oxide in porous residual fragments of rock. The striking appearance of the concretionary gravel in the Cr G horizon is due to its concentration owing to differential removal of other constituents by eluviation and weathering; and to the fact that it is more conspicuous, being harder, darker, and more thoroughly impregnated with iron oxide. The amounts of concretionary material in the coarse sand and gravel for a selection of samples analysed is shown in Table 1. On the average, in the layers between 6 and 48 in. in which concretions are formed and concentrated, the ratio of concretionary material to quartz is 2·4, 2·2, 0·9, 0·3, and 0·1 in the $>$ 7, 7–4, 4–2, 2–1, and 1–0·5 mm. fractions respectively. It thus appears that rock fragments smaller than 2 mm. do not as a rule harden to concretions.

The most immature concretions are soft, porous, of irregular outline, with a rough mat surface; they are red like the surrounding soil, or minutely mottled in red, yellowish-red, or brown. They grade into mature forms which are purple, very hard, dense (heavier than bromoform), smooth, and circular or elliptical. Flecks of mica or quartz grains are distributed haphazard throughout all types; and when the concretion is decomposed with hot dilute hydrochloric acid it is shown to contain much clay. Analyses of hydrochloric acid extracts showed that a batch of mature concretions contained 48 per cent. Fe_2O_3, and immature ones 33 per cent. There are ill-defined thin concentric layers of iron oxide on the outside of the mature concretion, but inside the iron oxide appears to have been deposited uniformly.

The increasing maturity of the residuals towards the surface is shown by the increasing percentage of hard purple concretions among the total number; namely 5, 10, and 30 per cent. in the 60–72, 36–42, and 6–12 in. samples respectively.

The explanation of the development of concretions such as these is still not entirely clear. Mobilization and relatively irreversible deposition of iron are essential. There is good evidence that iron may be mobilized even in a freely drained soil at a pH of about 6·5 by extracts from leaves (Bloomfield, 1951) or as a silica-protected sol (Demolon and Bastisse, 1944). This iron may be deposited either by evaporation during the dry

season or, should the iron be in the form of a ferrous complex, by oxidation, when a dry period may not be essential, for a more porous, better oxidized pocket in the soil would be sufficient. In this soil the deposition may occur in the porous fragments because there is probably a greater partial pressure of oxygen in the large pores; and also because there is more solution to be evaporated in a large pore so that the complex can become more concentrated. The large pore space will also lead to a more rapid drying and ageing of the deposit, and the better the deposit is aged the less likely it is to be peptized during the next rainy season.

Once the pores of the residual fragment have been coated with iron oxide, ferrous iron in solution as a complex may be absorbed on the surface of ferric oxide (Bloomfield, 1951). This may occur during a moist period, and then during a relatively dry period the ferrous iron will be oxidized and will build up the ferric oxide deposit within the pores.

These concretions do not, however, develop fully unless they are exposed to a succession of moist and very dry periods. A simple alternation of oxidizing and reducing conditions leads only to an iron oxide mottle. In this profile the hard concretions develop only in the upper part of the sedentary horizon, and it has frequently been noticed that this type of concretionary development ceases below the upper-slope member of the catena when subsoil conditions are moister; in fact, as will be shown later, these concretions begin to dissolve when they are carried by creep to the lower slope.

The surface deposit of iron oxide appears more stable the greater its radius of curvature. Hence very few concretions are found less than 1 mm. in diameter, and the concretions become more uniformly rounded as they develop. Possibly iron oxide is more easily mobilized the more curved the surface of the concretion, in the same way that the solubility of very small crystals is greater than large ones (Buckley, 1951). It is true that differences in the solubility of crystals larger than 1 μ are very slight indeed; but even slight differences in mobilization may prove significant over thousands of years.

D. *Free Iron Oxide in the Clay Fraction, and the Total Iron Oxide in the Soil*

In many granitic soils the red-coloured compact clayey subsoil contrasts sharply with the brown sandy top soil. The percentage of iron oxide in the clay fraction is nevertheless fairly constant in the top 42 in., and below that it declines gradually to the base of the oxidation-reduction zone at 19 ft., corresponding to the decreased weathering of the vermiculite (Table 1). There is a sharp drop in the reduced zone, partly because little iron from the vermiculite has been released, and also because most of the iron oxide in the reduced form will be removed in the drainage.

The total percentages of iron oxide in typical 2-mm. fine earth samples of soil increase to a depth of 6 ft., which may be related to the increasing proportion of relatively iron-rich clay in the soil. The amounts of iron oxide below that level are somewhat less, but there is still much iron in the pale-coloured reduced zone, and, as will be shown in the next part,

this is true for profiles farther down the slope. In many descriptions of deep profiles in the humid tropics it is often assumed without analytical evidence that the iron has been removed from the pallid to the mottled zone above by capillarity or by a fluctuating water-table. In the present profile most of the iron in the pallid reduced horizon is still combined as vermiculite and has not been lost. Any iron that has been released is probably removed downwards into the drainage. The red colour in the mottled zone is due to the weathering of the vermiculite in that zone, and the illuvial deposition of relatively iron-rich clay from above.

E. *Exchange Complex and pH*

There is a fairly steady pH of 6·5 to a depth of 6 ft.; it then falls to 5·5 and remains fairly steady until the reduced zone below 20 ft., where it rises sharply to 8·0. It is generally accepted that the higher pH on the surface is due to cations being brought up from the subsoil by the roots of plants and deposited on the surface in the plant remains. The depth at which the pH rises again has not been recorded before; and it may now be deduced that cations are largely removed by leaching or extraction by roots down to the capillary fringe.

The exchange capacity is consistent with the composition of the soil, and the percentage saturation follows the pH in a regular manner.

F. *The Textural Profile*

It is now possible to explain in detail the diverse series of textural horizons which are such a striking feature of this profile; and their discussion will serve to recapitulate the processes involved in its formation.

The Cr W, worm-formed, horizon. Mechanical analyses show a complete absence of coarse sand in the worm-casts and a greatly reduced amount in the 0–1-in. sample. The fine sand, silt, and clay occur in the same relative proportions as they do in the top 12 in. of the profile, so that the worms do not appear to ingest the mineral soil selectively below 0·2 mm.

The Cr T, termite-formed, horizon. This is almost gravel-free, in sharp contrast to the Cr G horizon below. In order to elucidate its development its composition is compared in Table 4 with that of the gravel and sedentary horizons below, on the assumption that no change in the 0·5–0·2 mm. medium sand fraction should have occurred. This is a reasonable assumption as the fraction is almost entirely quartz, and it is small enough to be moved by nearly all the ants and termites, together with the finer fractions. It is clear that there has been (*a*) a considerable eluviation of clay; (*b*) a loss of fine sand and silt which may be attributed largely to the weathering of the mica minerals; and (*c*) a progressive decrease in the 0·5–1, 1–2, and 2–4 mm. fractions due largely to the fact that many of the ants and termites are unable to bring up these larger fractions, and possibly to some physical comminution of the grains.

The Cr G horizon. The very high proportion of gravel and the rapidly increasing proportion of clay are the notable features. The average proportion of coarse quartz sand in the gravel horizon (6–26 in.) is about

TABLE 4

Mechanical Composition of the Upper Horizons relative to the 0·5–0·2 mm. Fraction (10)

Horizon	Depth (in.)	Size fraction (mm.)						
		4–2	2–1	1–0·5	0·5–0·2	0·2–0·02	0·02–0·002	Less than 0·002
Cr T .	2–3	2	8	6	10	3	1	3
Cr G_1 .	6–12	7	9	6	10	7	2	4
Cr G_2 .	18–24	16	16	7	10	7	5	24
S_1 .	36–42	17	19	8	10	10	6	42

double that in the sedentary horizon (26–120 in.). The amount of quartz gravel over $\frac{1}{2}$ in. (4·6 per cent.) is also nearly double the amount of quartz in veins in the sedentary horizon (2·5 per cent.). This was determined by measuring up with a ruler sample sections on the side of a nearby railway cutting. Thus the coarse sand and gravel represents the residue from a thickness of rock equal to about twice the thickness of the gravel horizon. This concentration is readily accounted for by weathering of the fine sand and silt fractions and eluviation of clay from the gravel horizon into the sedentary horizon.

The sedentary horizon. In this horizon there is a fairly constant proportion of coarse sand consisting of quartz and subsidiary felspar. The fine sand and silt fractions increase steadily with depth, for they contain a high proportion of mica which weathers in the upper levels. The clay content is at a maximum from 30 to 72 in., below which it gradually decreases due partly to the fact that less has been released by weathering, but mainly to the decreased illuviation below 6 ft. owing to scarcity of root channels. These changes continue to 20 ft., and below that the coarse sand begins to rise at the expense of the other fractions owing to the presence of compound fragments of decomposed rock.

It is necessary to emphasize that this profile, though mature, is developing continuously. The upper horizons are being removed by creep, clay is being eluviated through the fabric of decomposed rock, and at the base decomposed rock is emerging from the water-table. The clay is being eluviated down root channels from the base of the zone in which it is concentrated, and as it is removed more clay can take its place from higher up. There is no suggestion that clay is being washed through an already compact and static clay horizon.

Simonson (1950) considers that the processes of eluviation and illuviation are unable to account for the characteristic textural profiles of Red-Yellow Podzolic soils in the U.S.A. because the thick B horizons contain more clay than could have been derived from the relatively thin A horizons. If the idea of continuous development suggested for the present profile is correct, this objection does not apply in this case; for the large accumulation of clay between 30 and 72 in. has been derived not only by eluviation of the present upper layers but from former upper layers whose coarse constituents have been lost by creep. Simonson further suggests that the dominant processes involved in the development of the

marked textural profile are the formation of silicate clay minerals in the deeper horizons and the destruction of these minerals in the upper horizons. It does, however, seem very doubtful whether the clay in the upper horizons is destroyed. Pearson and Ensminger (1949) have examined the clay fractions of some typical Red-Yellow Podzolic soils and shown that the percentage of free iron oxide in the clay fraction is effectively constant down the profiles, though there are great differences in the amounts of clay in the A and B horizons; a strong argument in favour of translocation rather than destruction of clay. Furthermore, Scrivenor (1931), Vine (1949), and Waegemans (1951) have all given evidence that kaolin is very resistant to weathering under moderately acid free-draining conditions, so that it is unlikely that sufficient clay is destroyed to account for the very low amounts in the A horizon.

Summary

A detailed study has been made of all stages in the development of a deep profile formed over biotite gneiss in the West African forest. The rock had altered considerably below the water-table at 28 ft., mainly by decomposition of the felspars. In the zone of reduction above the water-table, the decomposed rock was very porous; and higher up, in the oxidation-reduction zone, clay was illuviated down root channels into the porous fragments of decomposed gneiss. These formed the nuclei of iron concretions.

Of the minerals in the gneiss, the felspars were weathered considerably below the water-table, but above it there was only slight weathering. Biotite was converted to vermiculite below the water-table, probably through a chlorite stage. In the oxidation-reduction zone the vermiculite decomposed farther, and iron oxide was deposited between the layers. Quartz grains tended to break into smaller fragments, and their sharp edges became rounded.

The clay mineral fraction consisted mainly of kaolinite with subsidiary goethite, a little quartz, and a trioctahedral hydromica. Some montmorillonoid was present near the water-table.

The iron-oxide concretions develop in the upper part of the sedentary horizon, although they harden and accumulate in the gravel horizon above.

In spite of very considerable eluviation of clay there is no differential movement of iron oxide in this fraction. There is no evidence that iron has been dissolved in the reduced zone and deposited in the oxidized zone above.

The usual fall in pH down the profile was observed. The point at which the pH rose again coincided roughly with the top of the capillary fringe.

The quartz gravel, accumulated in the gravel horizon, represents the residue from about twice the corresponding thickness of rock.

The profile is considered to be mature; but it is in a state of continuous development, whereby the top soil is being lost by creep, clay is being washed downwards through a fabric of porous decomposed rock, and, at the base, decomposed rock is emerging above the water-table.

Acknowledgements

It is a pleasure to acknowledge the help given me by Mr. J. C. Fatemehim and Mr. S. A. Akinleye in the work of analysis. For the X-ray diffraction determinations I have to thank Dr. A. Muir for so kindly allowing me the facilities of the Pedology Department, Rothamsted Experimental Station, and Mr. G. Brown for his indispensable guidance. I wish also to thank the Director of the Department of Geological Survey, Nigeria, for his help in the preparation of rock slides.

REFERENCES

BLOOMFIELD, C. 1951. Experiments on the mechanism of gley formation. J. Soil Sci. **2,** 196–211.

BROCK, R. W. 1943. Weathering of igneous rocks near Hong Kong. Geol. Soc. Amer. Bull. **54,** 717–38.

BROWN, G. 1951. X-ray Identification and Crystal Structures of Clay Minerals. Mineralogical Society, London, ch. 5.

BUCKLEY, H. E. 1951. Crystal Growth. Chapman & Hall, London.

CAMPBELL, J. M. 1917. Laterite—its origin, structure and minerals. Mining Mag. **17,** 67–77.

DEMOLON, A., and BASTISSE, E. 1944. Rôle vecteur de la silice dans les phénomènes géochemiques et physiologiques. Ann. Agron. **14,** 265–96.

DOYNE, H. C., and WATSON, W. A. 1933. Soil formation in Southern Nigeria. The Ilepa profile. J. Agric. Sci. **23,** 208–15.

GOLDICH, S. S. 1938. A study of rock weathering. J. Geol. **46,** 17–58.

MACEWAN, D. M. C. 1946. The identification and estimation of the montmorillonite group of minerals, with special reference to soil clays. J. Soc. Chem. Ind. **65,** 298–305.

MARSHALL, C. E., and JEFFRIES, C. O. 1946. Mineralogical methods in soil research. Soil Sci. Soc. Amer. Proc. (1945), **10,** 397–405.

MEINZER, O. E. 1923. The occurrence of ground water in the United States. U.S. Geol. Survey, Water Supply Paper 489.

MERRILL, G. P. 1897. Rocks, Rock-weathering and Soils. Macmillan, New York.

NYE, P. H. 1954. Some soil-forming processes in the humid tropics, I. A field study of a catena in the West African forest. J. Soil Sci. **5,** 7–21

PEARSON, R. W., and ENSMINGER, L. E. 1949. Types of clay minerals in Alabama soils. Soil Sci. Soc. Amer. Proc. (1948), **13,** 153–6.

PEECH, M. 1945. Determination of exchangeable cations and exchange capacity of soils—rapid micro-methods utilizing centrifuge and spectro-photometer. Soil Sci. **59,** 25–38.

SCHOFIELD, R. K. 1933. Rapid methods of examining soils—II. The use of para-nitrophenol for assessing lime status. J. Agric. Sci. **23,** 252–4.

SCRIVENOR, J. B. 1931. The Geology of Malaya, 143–4. London.

STEPHEN, I. 1952. A study of rock weathering, Part I. Weathering of biotitite and granite. J. Soil Sci. **3,** 20–33.

TATTAM, C. M. 1943. Groundwater. Farm & Forest, **4,** 71–75.

VINE, H. 1949. Nigerian soils in relation to parent materials. Comm. Bur. Soil Sci. Tech. Commun. No. 46, 22–29.

WAEGEMANS, G. 1951. Introduction à l'étude de la latérisation et des latérites du Centre Africain, Bull. Agric. Congo Belge, **42,** 13–56.

WALKER, G. F. 1949. Biotite weathering. Min. Mag. **28,** 693–703.

—— Trioctahedral minerals in the soil clays of north-east Scotland. Ibid. **29,** 72–83.

WALKLEY, A. 1947. A critical examination of a rapid method for determining organic carbon in soils. Soil Sci. **63,** 251–64.

(Received 22 May 1953)

SOME SOIL-FORMING PROCESSES IN THE HUMID TROPICS

III. LABORATORY STUDIES ON THE DEVELOPMENT OF A TYPICAL CATENA OVER GRANITIC GNEISS

P. H. NYE
(*University College of the Gold Coast*)

A COMPLETE field description of the members of a catena developed over a granitic gneiss at Ibadan in the forest region of south-west Nigeria has been given in Part I (Nye, 1954); and the soil-forming processes revealed by this field study were described. The details of the development of the upper slope profile have been examined in Part II (Nye, 1955). The present paper gives the results of laboratory studies on the remaining members of the catena, and with this knowledge the development of the whole catena is further examined.

Experimental

The following determinations were carried out on samples collected from the profiles described in Part I: mechanical analysis, exchangeable cations, exchangeable acidity, pH, organic carbon, and total iron on the whole soil; free iron oxide and estimation of minerals in the clay fraction; and identification of minerals in the fine and medium sand fractions. The frequency of sampling and methods adopted for the analysis have been described in Part II.

Discussion of Experimental Results

1. *Mechanical composition*

A selection of the mechanical analyses of the profile samples is given in Table 1, which confirms many of the field observations made on the characteristic horizons in Part I. The worm-formed horizon (Cr W), only about 1 in. thick, has a fairly constant composition in the upper slope and the upper and lower mid-slope profiles. In the other profiles it has been disturbed by recent cultivation. The termite-formed horizon (Cr T) deepens from 4 in. in the immature profile to about 24 in. in the upper mid-slope profile, and thereafter the thickness remains steady. Its composition is again fairly constant, and its low silt and clay content will be noted. In the gravel accumulation horizon (Cr G) there is an interesting variation in the distribution of the quartz gravel. Up-slope of the lower mid-slope profile, where the Cr G horizon is only some 2 ft. thick, there is no marked difference in the proportion of gravel throughout the horizon. But in the lower mid-slope profile and below it, where the Cr G horizon is some 6 ft. thick, there is a pronounced band of gravel about 1 ft. thick at the base of the horizon. It is possible that this band forms as a result of gradual settling. It is also possible that the

TABLE 1

Selection of Analytical Data (samples taken from the centre of the ranges)

	Upper slope				*Upper mid-slope*				*Lower mid-slope*					*Lower slope*						*Valley*				
Horizon	*Cr W*	*Cr T*	*Cr G*	S_1	*Cr W*	*Cr T*	*Cr G*	S_1	*Cr W*	*Cr T*	*Cr G*$_1$	*Cr G*$_2$	S_2	*Cr W and Cr T*	*Cr G*$_1$	*Cr G*$_2$	*Cr G*$_3$	*Cr G*$_4$	S_1	S_2	*Cr W*	*Cr G*$_2$	*Cr G*$_3$	*Cr G*$_5$	S_1
Depth (in.)	0–1	1–10	10–27	27–72	0–1	1–24	24–48	48–96	0–1	1–18	18–72	72–94	94–108+	0–18	18–30	30–48	48–84	84–90	90–102	102–8	0–24	24–48	48–96	96–108	108–32+
Colour	*dk. reddish-brown*	*reddish-brown*	*reddish-brown*	*red mottled*	*dk. reddish-brown*	*dk. reddish-brown*	*red*	*yellowish-red mottled*	*dk. greyish-brown*	*reddish-brown*	*reddish-brown*	*brown mottled*	*pale yellow*	*v. dk. grey-brown*	*strong brown*	*PAN*	*strong brown mottled*	*light brown-grey*	*light olive-grey*	*decomp. rock*	*grey-brown*	*light yellow-brown mottled*	*blue-green mottled*	*blue-green mottled*	*decomp. rock*
Sizes (mm.)																									
> 4%	..	..	15	2	..	..	15	17	..	..	11	30	1	..	21	..	15	35	1	52	..	1	..	6	2
4–0·5%	12	54	52	36	9	42	52	40	13	51	49	48	8	34	45	..	49	44	7	36	34	41	24	23	48
0·5–0·2%	21	21	9	6	19	31	15	11	17	27	10	6	3	25	22	..	10	6	7	20	25	30	22	17	22
0·2–0·02%	43	15	7	19	46	16	9	21	49	14	15	14	18	27	20	..	20	12	16	24	28	17	20	31	23
0·02–0·002%	11	4	4	10	11	5	4	10	12	3	6	8	27	7	6	..	7	8	14	11	8	5	7	15	4
< 0·002%	13	6	28	29	15	6	20	18	9	5	20	24	44	7	7	..	14	30	56	9	5	7	27	14	3
pH	7·2	7·2	6·3	6·4	6·8	5·9	6·1	6·0	7·0	6·7	6·2	6·5	5·4	6·5	6·3	..	6·4	6·3	5·7	6·7	5·9	6·5	7·2	7·1	7·1
Ex. cations (m.e./100 g.)																									
Ca	4·5	2·7	3·7	4·1	7·2	1·7	2·4	1·7	4·3	1·7	2·2	2·8	13·5	2·9	n.d.	..	2·5	6·2	13·1	n.d.	0·8	n.d.	5·5	10·2	10·3
Mg	1·3	0·95	1·16	1·22	2·23	0·64	1·34	1·08	1·52	0·97	1·32	1·55	3·41	1·13	n.d.	..	1·64	4·0	4·5	n.d.	0·53	n.d.	3·7	3·5	3·0
K	0·23	0·21	0·23	0·30	0·39	0·32	0·30	0·32	0·39	0·26	0·26	0·29	0·23	0·32	n.d.	..	0·30	0·29	0·31	n.d.	0·34	n.d.	0·32	0·36	0·32
Na	0·25	0·38	0·39	0·31	0·46	0·29	0·27	0·30	0·25	0·29	0·31	0·32	0·37	0·45	n.d.	..	0·35	0·34	0·63	n.d.	0·38	n.d.	0·52	0·33	0·40
Acidity	1·36	0·54	2·04	1·70	1·77	1·36	2·15	2·17	1·50	1·22	2·17	1·77	2·18	0·82	n.d.	..	1·22	2·04	2·44	n.d.	0·75	n.d.	0·41	0·82	0·82
Org. C%	2·58	0·52	0·25	0·12	2·46	0·24	0·14	0·11	1·09	0·33	0·18	0·10	tr.	0·60	0·12	..	tr.	tr.	tr.	tr.	0·22	tr.	tr.	tr.	tr.
Total Fe_2O_3%	2·4	1·9	4·8	6·3	2·2	1·7	3·9	6·6	2·0	2·0	5·1	7·6	6·9	2·1	2·2	..	6·2	6·9	5·9	5·4	1·5	2·6	3·4	n.d.	5·4
Free Fe_2O_3 in clay %	6·6	7·2	6·7	7·7	6·9	6·9	7·6	9·7	6·9	7·7	8·3	9·6	2·6	6·2	6·9	..	n.d.	7·9	1·7	n.d.	4·3	6·7	1·9	2·0	1·7

gravel band contains the gravel only recently brought into the creep horizon from points a comparatively short distance up-slope, the gravel from farther up having been reduced in size by abrasion or, more likely, fragmentation.

The depth to which clay has been eluviated increases steadily down the slope. For instance, the point at which the clay percentage rises above 10 per cent. is at 10 in. in the upper slope profile and 36 in. in the valley profile. The profile at the summit is too immature to show the pronounced eluviation found in the profiles below.

In the three lowest profiles there is a further sharp increase in the clay content at the water-table. In the lower mid-slope profile the increase appears at 94 in. in the upper part of the S horizon, but in the valley profile it occurs at 48 in. well up in the Cr G horizon. It seems most likely that the clay has been washed vertically down these profiles, and the eluviation has been arrested at the water-table where the laterally moving water is unable to remove it sufficiently fast. This accumulation of illuvial clay at the water-table was absent in the deep upper-slope profile described in Part II; and this may be explained by the observation that the lowest depth to which clay was being washed by vertical eluviation was only 16 ft. while the water-table was at 28 ft. It is unlikely that the clay is deposited in the lower profiles by the laterally flowing underground water, since there is no accumulation in the deep upper slope profile, and the water would be expected to remove rather than deposit clay.

2. *Estimation of the clay minerals*

The minerals identified in these profiles were the same as those found in the deep upper-slope profile, namely kaolin, montmorillonoid, a trioctahedral mica, quartz, and goethite. No new minerals were detected. The relative proportions between the mica, kaolin, and montmorillonoid deduced from the diffraction photographs of orientated aggregates are shown in Table 2.

Above the zone of reduction there is no montmorillonoid in any profile, and the proportion of kaolin to mica is about 10:1. In the zone of reduction high proportions of montmorillonoid appear in all profiles. It may be noted that in the valley the montmorillonoid appears high in the creep horizon, and has not been derived from decomposition of the minerals in the rock below. It is tempting to regard this montmorillonoid as being formed by the action of ground water rich in silica on kaolin. But this transformation has never been convincingly demonstrated in nature, and it is possible that the montmorillonoid is synthesized afresh from SiO_2, Al_2O_3, Mg, &c., in the ground water. In favour of the silication of kaolin is the apparent loss in the amount of kaolin in the creep horizon in the valley profile below the water-table. In the lower slope profile the creep horizon contains 15 per cent. clay, i.e. 13 per cent. kaolin. The creep horizon in the valley-profile below the water-table has 27 per cent. clay, i.e. 6 per cent. kaolin only, in spite of the increased clay content.

A powder diagram of the clay from the base of the seepage iron pan

TABLE 2

Percentages of Mica, Kaolin, and Montmorillonoid in the Clay (< 2 μ) Fraction*

Horizon	*Depth (in.)*	*Mica*	*Kaolin*	*Montmorillonoid*
		Upper Mid-slope profile		
Cr T_2	12–24	9	91	..
Cr G	24–48	9	91	..
S_1	48–96	3	97	..
		Lower mid-slope profile		
Cr G_1	18–72	12	88	..
Cr G_2	72–94	3	97	..
S_2	94–108	4	53	43
		Lower slope profile		
Cr G_3	48–84	9	91	..
Cr G_4	84–90	14	72	14
S_1	90–102	18	25	57
		Valley profile		
Cr G_1	24–36	8	84	8
Cr G_3	48–84	4	23	73
Cr G_3	84–96	6	29	65
S_1	108–132	13	13	74

* The accuracy of these figures is limited by the visual estimation of the intensity of the diffraction lines.

(48–60 in.) in the lower slope profile showed no evidence of lepidocrocite being formed by oxidation of the ferrous iron in the drainage water, though there was abundant goethite.

3. *The exchange complex*

A selection of the results of the measurements on the exchange complex—the exchangeable cations, exchangeable acidity, and pH are shown in Table 1.

pH. The behaviour of the pH in a catenary sequence has been studied by Morison (1948) in the southern Sudan. He found there was a fall in pH with depth on the upper slopes and a rise in the valleys; the intermediate colluvial soils behaved erratically. In this catena the immature profile shows no marked change in pH; the upper, and upper and lower mid-slope profiles show the usual increase in acidity with depth, which was discussed in Part II; the lower slope profile maintains a steady pH; and the valley profile shows a steady rise in pH from 5·5 at the surface to 7·2 at the water-table at 4 ft. The behaviour of the pH in the lower profiles probably depends largely on the type of drainage. If, as in these soils, the water percolates downwards to the water-table and then runs laterally, the subsoil is not being enriched with bases above the water-table, and the pH may fall with depth. If, however, there is lateral percolation of drainage through the subsoil so that the lower slope sub-

soils receive cations leached from higher up the slope, then the pH may be expected to rise.

The behaviour of the pH near the water-table in the lower mid-slope and lower slope profiles is interesting. On passing from the zone of oxidation-reduction to the reduced zone at about 94 in. there is first a sharp drop in pH together with a striking rise in the Ca and Mg contents of the exchange complex. About 1 ft. below there is a sharp rise in pH to near that of the ground water. The fall in pH and rise in the Ca and Mg content is associated with the appearance of montmorillonoid in the clay fraction.

Exchangeable cations. The amounts of Na and K remain fairly steady throughout the catena. The amounts of Ca and Mg are also fairly constant in the subsoils, which are kaolinitic and contain about 20 per cent. of clay; but they rise in the surface soil and near the water-table. In the surface soil the rise in the Ca and Mg contents corresponds to the increased amounts of organic matter, and near the water-table to the presence of montmorillonoid. The greatly increased ratio of divalent to monovalent cations when montmorillonoid is present is consistent with the exchange properties of montmorillonoid and kaolin clays (Russell, 1950).

4. *Free iron oxide in the clay fraction, and total iron oxide in the soil*

The percentage of free iron oxide in the clay fraction is fairly constant within each profile above the zone of reduction, and it also shows little variation between profiles. Thus the iron is showing little tendency to be differentially leached from the clay as it creeps downhill into the moister lower-slope profiles, although the colour change in the subsoil from red in the upper slope to strong brown in the lower slope indicates an increasing degree of hydration. This stability of the iron oxide in the clay fraction may be contrasted with that of the concretionary iron oxide, which tends to dissolve in the lower slope profiles (*v. infra*). In the zone of reduction, however, there is a sharp decrease in the amount of free iron in the clay, and even some of the iron found may have been released owing to attack on the montmorillonoid and mica clay minerals by the hydrosulphite.

The total amounts of iron oxide in the 2-mm. samples increase with depth in the upper, and upper and lower mid-slope profiles, owing to illuviation of the relatively iron-rich clay. The lower levels of the lower mid-slope horizon may have their iron content increased by deposition from seepage water, and this process is certainly responsible for the iron pan in the Cr G horizon of the lower slope profile. In the grey reduced sedentary horizons at the base of the three lowest profiles the content of iron oxide is still relatively high. This fact was noted in the pallid zone at the base of the deep upper slope profile, and may be attributed to the iron in the vermiculite (see Part II).

5. *The development and decay of iron and manganese concretions*

It has already been shown in Part II that pisolitic iron concretions are formed in the upper slope profile by deposition of iron oxide in small

decomposed rock fragments in the upper part of the sedentary horizon. They are then accumulated in the creep horizon. The formation of concretions in the profiles lower down the slope, and the changes which take place in concretions formed on the upper slope as they creep downhill, may now be considered.

Near the top of the upper slope profile 75 per cent. of the concretions show a polished yellow-brown or purple exterior and the remainder have a mat surface. In the upper mid-slope profile these concretions have lost their fresh hard-polished appearance and all have dull yellow-brown surfaces. In the same profile an entirely different kind of concretion appears. At the top of the S horizon, and in the Cr G horizon, there are a few soft black manganese oxide mottles up to $\frac{1}{2}$ in. in diameter, which are not deposited around any obvious nuclei. On some of them yellowish-red iron oxide is deposited which hardens towards the surface to give a large concretion.

In the lower mid-slope and lower slope profiles both types of concretion have a very weathered appearance, showing a dull brown mat surface with grains of quartz protruding, which indicates that they are being slowly dissolved. In the valley profile, the concretions are similar above the water-table, but below it the last stages of dissolution are shown, and the majority are nearly white. The soft manganese oxide mottles are present in the oxidation-reduction zone in each of these three profiles.

The conclusion that hard pisolitic iron concretions may dissolve slowly in the zone of oxidation is supported by the observations of Carroll and Jones (1947) in Australia and by Campbell (1917), who shows photo-micrographs of the partially dissolved pisolites.

Contemporary formation of iron concretions. Doubt is often expressed whether 'laterite' is being formed at the present time; e.g. the latest authorities, Prescott and Pendleton (1952), who use the word to include ferruginous gravels, write: 'It is further obvious that most, if not all occurrences of laterite can be traced to Tertiary times, . . .' They also suggest very tentatively that Doyne and Watson's (1933) Ilepa profile of south-west Nigeria, which is very similar to the upper slope profile described, 'can also possibly be interpreted as an undisturbed relict soil'. Vine (1949) also notes a progressive change from the red parts of the mottled clay to hard gravel and states that it is not clear how much of this is a current process.

The author is unable to accept this static view. It has been shown in Part I that the horizon of gravel accumulation is creeping slowly downhill, and, in the next part, it will be shown that the rate of lowering of the surface is of the order of at least 1 ft. in 12,000 years. Thus any relict laterite in the top 2 ft. would be removed by creep within some 25,000 years. There is also no laterite cap at the summit of the slope which could act as a continuous source of relict laterite creeping downhill. It seems clear, then, that the iron gravel in the Cr G horizon has been formed within the last 25,000 years. There is a general tendency towards a greater concentration of concretions and the formation of iron pan in the savannah, and a reduction in the amount of concretions towards the wetter parts of the forest (Vine, 1949), and it is possible that

climatic changes in the past have influenced the extent of the process, but that it is continuing today there can be little doubt.

6. *The development of the iron pan*

The formation of a seepage iron pan in the lower slope profile between 30–48 in. has already been described in Part I. The total iron content of the pan is 7·3 per cent. This comparatively small amount of iron oxide is able to cement the soil because it has been deposited and dehydrated to form a hard ferruginous framework within which the earth is much less ferruginous.

A point that has not been considered in Part I is the way the pan develops as the surface of the catena is lowered in the normal course of erosion. It is thought that in general the iron is dissolving at the top of the pan and being redeposited below, and in this way the development of the pan is keeping pace with the lowering of the surface. The appearance of the profile at the top of the pan certainly suggests it is dissolving; for, as will be recalled, the pan consists of iron encircled knobbles in a ferruginous matrix, whereas in the horizon immediately above it there is no ferruginous matrix and the abundant concretions have the appearance of weathering knobbles.

At the top of the pan there are two opposing tendencies operating. The relatively iron free drainage-water descending vertically is tending to dissolve the iron, as evidenced by the dissolving of the iron in concretions in the lower part of the catena; and the relatively iron-rich ground water, seeping laterally, is tending to deposit iron in the oxidized parts of the subsoil. As erosion proceeds, the valley deepens, and the level of the seepage water will be lowered. The top of the pan will thus be carried above the influence of the deposition process and will dissolve.

If the deepening of the valley is very rapid, the pan may not dissolve sufficiently fast, and will outcrop on the surface, forming a distinct ledge; though only a few, scattered outcrops occur in this region. Milne (1936) in East Africa has mentioned the emergence of a clinker-like horizon at the surface in the lower slope of a catena, which corresponds to the outcrop under discussion. Greene (1950) has described the contemporary formation of lateritic material in valley sites in the Southern Sudan, and pointed out the dangers in assuming such processes are irreversible.

7. *Mineralogical examination of the valley profile*

In Table 3 the mineralogical composition of the 0·5–0·2 mm. and the 0·2–0·02 mm. fractions of the valley profile are compared with those of the creep horizon of the upper slope profile to discover what further changes have occurred in the minerals as they creep downhill.

Mica. It will be noted that micaceous minerals have nearly disappeared from the creep horizons in the valley profiles; and the few flakes remaining are further altered. In the upper slope they were dark brown or black owing to deposited ferric oxide; in the valley they retain the dark coloured centre, but are often clear at the edges where the iron has dissolved. The refractive index at the edge is frequently lower than 1·54.

In the sedentary horizon there is an abrupt change, micaceous minerals

TABLE 3

Mineralogical Composition of the Valley Profile

Horizon	*Depth (in.)*	*Quartz*	*Felspar*	*Mica*	*Zircon*
	Medium sand (0·5–0·2 mm.)				
Cr T	6–24	87	13	..	..
Cr G_1	24–36	91	9	..	..
Cr G_2	48–84	93	7	..	..
Cr G_3	84–96	91	6	3	..
S_1	108–132	32	1	67	..
Cr G	Upper slope profile	94	1	4	..
	Fine sand (0·2–0·02 mm.)				
Cr T	6–24	47	50	3	..
Cr G_1	24–36	66	34	..	..
Cr G_2	48–84	50	38	10	2
Cr G_3	84–96	60	19	20	1
S_1	108–132	5	3	92	..
Cr G	Upper slope profile	53	20	26	1

being dominant in the fine sand, and abundant in the medium sand fractions. About one-third of the mica is of the olive chloritic form described in Part II, and the remainder is fresh yellow or golden vermiculite. The rock at this point probably contains a biotite-rich band.

Felspar. There is a surprisingly high proportion of felspar in both fractions. The proportions of potash felspar to plagioclase were approximately 10:1, a determination which was checked, in the absence of mica, by the staining technique described by Hawkins and Graham (1951). It was noted in Part II that in the deep upper-slope profile there was little change in the proportions of quartz and felspar to a depth of nearly 30 ft., and the high proportion of felspar in the valley confirms the remarkable stability of the felspar grains that survive the initial severe destruction below the water-table.

8. *Relationship between Ibadan profiles and Buchanan's laterite*

Laterite, both hard iron pan or concretions, and soft lithomarge (Fox, 1936) is usually described from areas with uplifted peneplains, originally of low relief associated with high water-tables (Prescott and Pendleton, 1952). Laterite on ancient peneplains is well developed in Northern Nigeria (Vine, 1949; du Preez, 1949). However, the descriptions and photographs of Fox show the topography of the Angadipuram area of Malabar, where Buchanan first described laterite, to be very similar to the rolling country, with well rounded divides, round Ibadan, south-west Nigeria. The photographs show that the laterite quarries are on the middle slopes of the hillsides, and their surface is at least 40 ft. below the top of the hill. They are thus too deep to have developed under formerly nearly level topography with a high water-table. The lithomarge is more likely similar in origin to the mottled illuvial clay subsoil found in the upper and mid-slope profiles developed over granitic gneiss in south-west Nigeria, examples of which have been described. The

quarried zone appears to correspond to the zone of maximum clay content and maximum density, both features likely to be important in brick-making. Good bricks cannot be made from the particular Ibadan profiles described here, probably because of the very high content of coarse quartz sand, but in areas nearby where there is less coarse quartz in the gneiss the lithomarge sets very hard on drying. The lithomarge associated with ancient peneplains may well have been formed under conditions different from those in Buchanan's type area.

Summary

An account is given of laboratory studies made on the catena developed over granitic gneiss in the West African forest, which was fully described in Part I.

1. The worm- and the termite-formed horizons have a fairly constant composition on all parts of the slope. The thickness of soil from which clay is eluviated increases steadily down the slope; and in the lower slope there is a further increase in the clay content at the water-table where the eluviation of clay is arrested.

2. The clay minerals in the free-draining portions of the profiles are predominantly kaolin, with subsidiary quartz, goethite, and a trioctahedral mica. In the reduced zones there is much montmorillonoid, which is probably formed by silication of kaolin.

3. The pH falls with depth in the upper and mid slopes, remains steady in the lower slopes, and rises in the valley. At the point where montmorillonoid is formed there is a sharp drop in pH and a rise in exchangeable Ca and Mg. The content of exchangeable cations remains fairly steady, except for the rise in Ca and Mg near the reduced zone and also in the humic top soil.

4. The free iron-oxide content of the clay fraction remains fairly steady in the zones of oxidation and oxidation-reduction throughout the whole catena. There is a sharp decrease in the reduced horizons.

5. Iron concretions, formed in the upper-slope profile, dissolve as they are carried by creep to the moister lower slopes. Another type of concretion is formed in the mid- and lower-slope profiles by deposition of iron oxide around soft manganese-oxide concretions. The formation and dissolution of concretions is a contemporary process.

6. It is suggested that the conditions for formation of the mottled subsoils in the upper- and mid-slope profiles closely resemble those responsible for the vermicular lithomarge described by Buchanan.

REFERENCES

Buchanan, F. 1807. A Journey from Madras through the Countries of Mysore, Canara and Malabar, etc. (in 1800–1). London.

Campbell, J. M. 1917. Laterite—its origin, structure and minerals. Mining Mag. **17**, 171–9.

Carroll, D., and Jones, N. K. 1947. Laterite developed on acid rocks in South-Western Australia. Soil Sci. **64**, 1–15.

Doyne, H. C., and Watson, W. A. 1933. Soil formation in Southern Nigeria. J. Agric. Sci. **23**, 208–15.

Du Preez, J. W. 1949. Laterite: a general discussion with a description of Nigerian occurrences. Bull. Agric. Congo Belge, **40**, 53–66.

Fox, C. S. 1936. Buchanan's laterite of Malabar and Kanara. Records of the Geological Survey of India, **69**, 389–422.

Greene, H. 1950. So-called irreversible laterisation. 4th Int. Cong. Soil Sci. Trans. **2**, 175–7.

Hawkins, R. H., and Graham, E. R. 1951. Mineral contents of the silt separates of some Missouri soils as these indicate the fertility level and degree of weathering. Soil Sci. Soc. Amer. Proc. (1950), **15**, 308–13.

Milne, G. 1936. Normal erosion as a factor in soil profile development. Nature, **138**, 548.

Morison, C. G. T., Hoyle, A. C., and Hope-Simpson, J. F. 1948. Tropical soil-vegetation catenas and mosaics. J. Ecol. **36**, 1–84.

Nye, P. H. 1954*a*. Some soil-forming processes in the humid tropics. I. A field study of a catena in the West African forest. J. Soil Sci. **5**, 7–22.

—— 1955. Some soil-forming processes in the humid tropics. II. The development of the upper-slope member of the catena. Ibid. **6**, 51.

Prescott, J. A., and Pendleton, R. L. 1952. Laterite and lateritic soils. Comm. Bur. Soil Sci. Tech. Commun. **47**.

Russell, E. W. 1950. Soil Conditions and Plant Growth. Longmans, London, p. 109.

Vine, H. 1949. Nigerian soils in relation to parent materials. Comm. Bur. Soil Sci. Tech. Commun. **46**, 22–29.

(*Received 22 May 1953*)

SOME SOIL-FORMING PROCESSES IN THE HUMID TROPICS

IV. THE ACTION OF THE SOIL FAUNA

P. H. NYE

(*University College of the Gold Coast*)

In Part I of this series (this Journal, **5**, p. 7) it was shown that the soil fauna played an important part in the formation of the upper layers of soils in the forest belt of south-west Nigeria. Above a horizon of gravel accumulation (Cr G horizon), there was shown to be a gravel-free coarse sandy horizon formed largely by termites (Cr T horizon), and above this a thin horizon of fine sandy loam formed by worms (Cr W horizon). These three horizons together form the creep (Cr) horizon in which the soil is moving downhill. The sedentary (S) horizon is below. In this paper the activities of the fauna in the area at Ibadan previously described will be discussed in some detail.

Termites and Ants

Review of literature. Few detailed studies of the influence of ants or termites on soil formation have been made, and these have usually concentrated on the chemical and mechanical composition of termite mounds rather than the role of termites in moving soil, which is the main object of the present study. Long ago, Shaler (1891) observed that on a 4-acre field at Cambridge, Mass., ants transferred at least $\frac{1}{5}$ in. of soil to the surface each year during 2 years' observation, and he refers to a field in New England 'covered to a depth of some inches by a fine sand' clearly deposited by ants. In neither case was the depth from which the material was obtained quoted. There are, indeed, only a few observations recorded about the working depth of ants and termites. Kalshoven (1941) states that termites in the East Indies seek the level of the ground-water at depths of at least 4 metres; and Marais (1950) has observed at Waterberg, South Africa, a gallery, leading to a well shaft, which extended 40 ft. vertically and 30 ft. horizontally from the termitarium. Harris (1949) in Uganda has observed that *Bellicositermes natalensis*, which is closely related to *Macrotermes nigeriensis* discussed in this paper, builds its mounds rapidly with soil collected down to 6 or 9 ft.; small-mound builders, on the other hand, work almost entirely in the topsoil.

The feeding habits of termites, their building materials, and their influence on soil fertility have been reviewed by Adamson (1943), who stressed the need for more detailed study and records. There is no general agreement about the difference between the chemical composition of the mounds and the surrounding soil, which seems to vary according to the species of termite, the age of the mound and the part of it sampled, and the type of soil. These important facts have rarely been

mentioned. A general account of termites, which indicates their extraordinary abundance and diversity in the tropics, has been given by Hegh (1922).

In the area around Ibadan a useful guide to the depth of origin of the insect deposits is their colour; earth derived from above 12 in. being predominantly brown, and below that red. Interest was concentrated on *Macrotermes nigeriensis* Sjost.,* for this is the species most active in bringing up earth from a depth.

The activities of Macrotermes nigeriensis *Sjost*

The development and construction of the termitaria of *Bellicositermes natalensis* have been described from an entomological standpoint by Hegh (1922), who shows many photographs and gives abundant references to their occurrence all over tropical Africa. The mounds of *Macrotermes nigeriensis* appear to be very similar. Two were investigated in detail. Mound A was only 10 yds. from the deep upper slope profile described in Part II (this Journal, 51). It had been fairly recently abandoned. Mound B, which was active, was on a similar site near Ijebu-Igbu, 30 miles south of Ibadan, which had been freshly cleared from high forest. Half of each mound was excavated, and a pit dug below it, to reveal a vertical section.

Description of mounds. A scale diagram of Mound A showing the relation of its galleries to the different horizons of the soil appears in Fig. 1. Samples were taken of the casing and nest portions of the mounds, and of the soil horizons below them. The mechanical and chemical compositions of these samples are compared with those of undisturbed profiles nearby in Tables 1 and 2 respectively.

The casing of Mound A is a compact red coarse sandy clay which differs little from the outside to the inside or from the summit to the base. It is composed of a mass of pellets about 1–2 mm. in diameter. The method of construction may be observed if the casing of a live mound is broken, when, after a few minutes, the termites will be seen bringing up the moistened pellets in their mandibles and carefully placing them in position to repair the break. The casing is full of channels, circular or elliptical in section and up to 9 in. across. It will be seen from Fig. 1 that the casing is thin at the top, and that rain wash has carried earth from the mound a distance of at least 3 ft. from the base. In young developing mounds the nest is not separated from the casing, but in mature mounds it is attached to the casing at a few points only.

The framework of the nest is composed of a compact mass of pellets like the casing, but lacks the coarse sand fraction. It forms large cells some 2–4 in. in diameter which are filled with white cellular fungus gardens. The red framework is often stained black by fungus bodies, probably secondary.

Special interest attaches to the galleries which radiate from the floor of the nest, for they give an indication of the depth to which the soil has been disturbed in making the mound. Immediately below the nest, to a

* Kindly identified by Professor J. E. Webb.

depth of about 36 in., the soil is full of galleries about $\frac{1}{4}$–$\frac{1}{2}$ in. wide, stained black on their sides. At irregular intervals they open into chambers up to 3 in. wide. From this thoroughly disturbed region below the mound

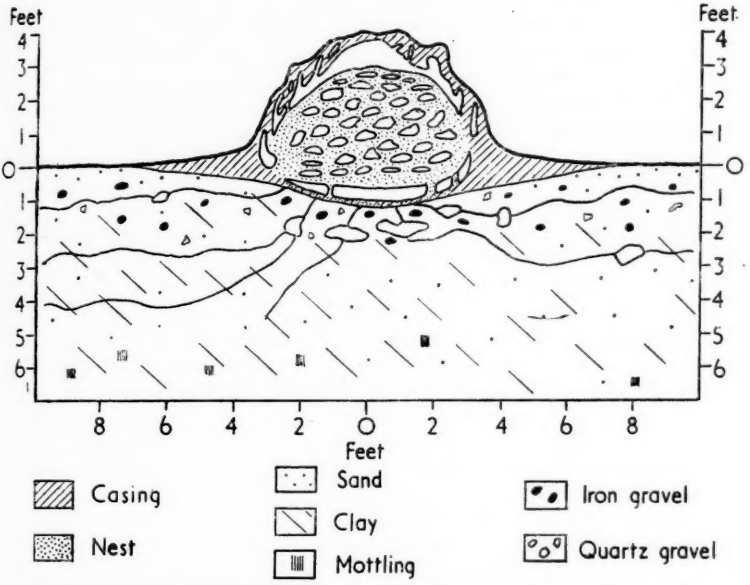


FIG. 1. Relation between mound of *Macrotermes nigeriensis* and the underlying soil.

the galleries extend, in the main, horizontally; and from the appearance of the side of the pit 6 ft. from the foot of the mound, the main working depth of the termites appeared to be from 6 to 48 in. Below 48 in. chambers were rare and galleries not easy to detect. Under the living Mound B, however, it was easier to pick out the galleries, for they were filled with greyish-brown frass. Here a thorough search below the mound revealed only two galleries at 10 ft., rather more at 6 ft., and many above 4 ft.

Mechanical composition of the mounds. It will be seen from Table 1 that there is an upper limit to the size of particles in the casing at 4 mm., and a relatively low proportion between 2 and 4 mm.; below 2 mm. the proportions in the various size fractions are similar to those in the 12–30-in. layers of the soil. The termites appear, therefore, to be reluctant to carry particles larger than 2 mm. and unable to carry them larger than 4 mm. In the nest, however, there is a sharp drop in the proportion of particles above 0·5 mm. To account for this it is suggested that the nest, in contrast to the casing, is built mainly from earth that has passed through the bodies of the termites, they being unable to pass grains larger than 0·5 mm. When sections of the gut contents of three major

TABLE 1

Mechanical Composition of Termite Mounds compared with Undisturbed Profiles nearby

Sample	Size fraction (mm.)								
	>7	7–4	4–2	2–1	1–0·5	0·5–0·2	0·2–0·02	0·02–0·002	<0·002
Mound A									
Casing, top	..	..	2	22	12	12	11	8	33
,, near ground	..	..	5	18	12	15	13	9	28
Nest, upper part	..	..	..	1	4	11	17	9	58
,, lower part	..	..	..	1	3	10	15	14	57
Below mound 6–12 in.	37	19	21	12	7	12	12	7	29
,, ,, 12–30 in.	8	14	30	17	8	8	7	5	25
Undisturbed Cr T 1–6 in.	..	2	7	23	19	26	13	5	7
Profile Cr G_1 6–12 in.	26	10	16	19	14	22	14	6	9
,, Cr G_2 12–30 in.	3	11	19	19	8	12	8	6	28
Mound B									
Casing, top	..	..	1	6	6	16	16	7	48
,, near ground	..	..	1	4	5	21	18	8	43
Nest, upper part	..	..	..	..	1	13	15	9	62
,, lower part	..	..	..	..	2	14	15	10	59
Below mound 11–30 in.	29	10	10	6	4	16	16	7	41
Undisturbed Cr T 0–5 in.	1	..	1	4	7	36	19	6	27
Profile Cr G_1 5–12 in.	37	11	13	7	6	23	16	5	30
,, Cr G_2 12–30 in.	8	6	6	5	5	14	20	12	38

TABLE 2

pH, *Exchangeable Cations, Exchangeable Acidity, and Organic Carbon in Termite Mounds compared with Undisturbed Profiles nearby*

Sample	pH	Exchangeable cations and acidity (m.e. per 100 g.)						Sat. %	Org. C. %
		Ca	Mg	K	Na	Mn	Acidity		
Mound A									
Casing, top	6·7	3·2	1·8	0·43	0·40	0·04	1·5	80	0·79
,, near ground	6·5	3·4	1·6	0·40	0·44	0·05	2·3	72	0·84
Nest, upper part	6·7	4·9	2·0	0·42	0·42	0·09	2·3	72	0·91
,, lower part	7·4	4·1	1·4	0·41	0·42	0·11	0·4	94	1·50
Below mound 6–12 in.	5·7	4·1	1·5	0·45	0·34	0·11	2·5	72	1·42
,, ,, 12–30 in.	5·4	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0·37
Undisturbed Cr T 1–6 in.	6·4	2·5	1·3	0·43	0·49	n.d.	2·0	70	0·53
Profile Cr G_1 6–12 in.	6·7	1·9	0·8	0·34	0·50	n.d.	1·6	69	0·32
,, Cr G_2 12–30 in.	6·7	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0·30
Mound B									
Casing, top	6·8	3·8	0·9	0·41	0·35	n.d.	2·0	74	0·58
,, near ground	6·4	3·2	0·7	0·32	0·31	n.d.	2·1	68	0·65
Nest, upper part	7·0	4·0	1·3	0·41	0·38	n.d.	1·6	83	0·91
,, lower part	7·1	6·2	1·0	0·34	0·21	n.d.	2·0	80	0·70
Below mound 11–30 in.	7·0	5·6	0·8	0·34	0·33	n.d.	2·0	78	0·41
,, ,, 48–72 in.	6·6	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
,, ,, 96–120 in.	5·5	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Undisturbed Cr T 0–5 in.	6·5	4·6	1·1	0·39	0·38	0·04	2·4	74	1·70
Profile Cr G_1 5–12 in.	6·1	2·5	0·8	0·30	0·38	n.d.	2·3	64	0·70
,, Cr G_2 12–30 in.	6·5	3·6	1·0	0·38	0·35	n.d.	2·0	73	0·25
,, S 48–72 in.	5·1	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
,, S 96–120 in.	5·2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

workers and three major soldiers were examined microscopically,* the maximum size of grains found was 0·15 mm.

Careful matching of the colour of the casing against the different soil horizons beneath the mound suggests the origin of the material is on the average between 12 and 18 in. deep; while the mechanical composition of the casings, as noted above, indicates that the earth is mainly taken from between depths of 12 and 30 in. Except near the base of the mound few of the pellets which form the casing are coloured brown, suggesting that the termites do not use the humic surface 6 in. as building material.

Chemical composition of the mounds. The organic carbon, pH, and base status of the mounds, and the soil below them, are compared with the values for the soil nearby in Table 2. In both mounds the casings vary little in composition from the summit to the base. Their organic-carbon contents are somewhat higher than those of the earth from which they have been formed, due possibly to the secretions used to stick the earth pellets together. In other respects their composition differs little from the original earth.

The samples of the nest framework and the soil layers below contain varying amounts of termite excretion and fungal matter. The organic-carbon content of the nest is naturally higher than in adjacent profiles; the pH, exchangeable Ca and Mg, and saturation percentage are also higher. The pH in Mound A drops rather sharply below the mound, owing possibly to acid decomposition of the fungal matter, but in Mound B it remains well above that of the nearby soil to a depth of at least 6 ft. There was no sign at the base of these mounds of an accumulation of calcium carbonate which has been found by many observers in larger and denser dome-shaped mounds (Griffith, 1938; Pendleton, 1942; Milne, 1947); although in the dry southern part of the Gold Coast under a rainfall of about 30 in. calcium carbonate is accumulated at the base of mounds similar to those described here.

On the vexed question of the relative fertility of termite mounds compared with the surrounding soil it can be said that the standing mound does not support vegetation, probably because the casing is too compact and dries out too readily; and that the collapsed mound has a topsoil poorer in nutrients and more compact than the surrounding topsoil, but below the mound the subsoil may be richer. Annual crops such as maize grow relatively poorly on levelled-off or collapsed mounds. These observations agree with those of Griffith (1938) in Uganda on the conical mounds of red earth that appear to be similar to those described here. However, all these effects are of minor account compared with the importance of the termites in forming a gravel-free topsoil. It is likely also that in these frost-free areas the fauna is the main agent disturbing the soil, and thus accelerates creep. This process, by continually lowering the topsoil, helps to prevent the soil from becoming too deeply weathered, and enables the roots of trees to keep in touch with decomposing minerals releasing nutrients in the subsoil.

The growth and decay of the mounds, and the rate of soil formation. The mounds are built very rapidly. A mound 2 ft. high has been seen to

* The sections were kindly provided by Mrs. J. E. Webb.

build in a month, and mature mounds 5 ft. high often have annual herbs, which are rooted in the soil surface, projecting from the casing. The termites do not appear to live long in the mature mound, for the proportion of living to abandoned mounds was 1:15. The mature mound collapses slowly after it has been abandoned. The top is reduced by rain wash and the upper parts of the walls collapse inwards. The holes up to 1 ft. in diameter at the base of mature mounds leading into the central chamber are not, as was first thought, due to burrowing animals, but to the washing away of the outer walls to expose the tunnels within the casing. It is thought that the mounds grow and decay to half their maximum height in 5 to 10 years. This estimate is based on the observation of a reliable authority on a mound in his garden. In addition, mounds are often built around trees and shrubs, and young trees less than 10 years old may sometimes be seen growing out of mounds which have broken down to less than half their original height.

Mound A was of average size for a mature mound in this area. It was found by direct weighing to contain 2·5 tons of earth (casing 3,970 lb., nest 1,570 lb.). The frequency of mounds is very variable. In the Ibadan region there were 2 mounds per acre above half-height; but in the freshly cleared forest at Ijebu-Igbu there were 25 such mounds per acre, though general observation suggests this is an exceptionally high density.

From these figures an idea of the rate of soil formation and the rate of lowering of the land surface by creep may be deduced. On the Ibadan site it may be calculated that at least $\frac{1}{2}$ ton of earth per acre is being deposited on the topsoil by termites every year. From the depth to which the termites are excavating it may be deduced that on the upper slopes about one-third of this material will be derived from the sedentary soil below the creep horizon. Therefore at least 2,000 tons per acre, or 1 ft. of soil, will be transferred to the creep horizon every 12,000 years. With a greater concentration of mounds this rate may be considerably increased.

The figure of 1 ft. per 12,000 years also represents the minimum rate at which the land surface is being lowered, since if equilibrium conditions are established, the rate of addition of materials to the creep horizon will equal the rate of loss by creep. It is interesting to compare this estimate with the rates of denudation estimated for other areas. Ashley (1930) has estimated that the hardest rocks forming the highest ridges of the folded Appalachian mountains are being lowered by 1 ft. in 10,000 years; and from the amount of sediment carried by the Mississippi river, quoted by Clarke (1924), it may be calculated that the surface of the basin is being lowered at the rate of about 1 ft. in 4,000 years (370,000,000 metric tons per annum from a drainage basin of 1,265,000 sq. miles).

Other ants and termites

The activities of other termites and ants which are conspicuously active in the Ibadan area may also be mentioned.

Earth deposited by termites covering fallen woody material is seen everywhere. It is often left as a tunnel when all the wood inside has been consumed. It is sometimes a brown and sometimes a red fine sandy

loam. The mechanical analysis of the red earth is shown in Table 3. The ratio of clay to silt and fine sand shows that it has been derived from the top 12 in. of soil.

TABLE 3

Mechanical Composition of Termite, Ant, and Worm Deposits

Sample	*Size fraction (mm.)*								*Clay ÷ (silt + f. sand)*
	7–4	*4–2*	*2–1*	*1–0·5*	*0·5–0·2*	*0·2–0·02*	*0·02–0·002*	*<0·002*	
Red earth covering twigs	..	..	4	8	28	35	10	15	0·33
,, ,, from large ant	..	5	22	18	32	14	4	5	0·28
Brown earth from small ant	..	..	10	7	18	42	9	13	0·26
Worm casts	..	..	..	1	15	45	14	25	0·42
Undisturbed upper slope profile:									
Cr W 0–1 in.	1	3	12	10	29	26	7	13	0·39
Cr T 1–6 in.	2	7	23	19	26	13	5	7	0·39
Cr G_1 6–12 in.	10	16	19	14	22	14	6	9	0·45
Cr G_2 12–18 in.	13	17	21	13	17	10	5	17	1·13
Cr G_2 18–24 in.	11	19	19	8	12	8	6	28	2·00

Hemispherical mounds, about 1 ft. high, of *Nasutitermes* sp. are common, but the brown colour of the earth shows that the soil is derived from no great depth. Harris (1949) in Uganda has noted that small-mound builders work almost entirely in the topsoil.

Black ants of all sizes were active in depositing loose sandy loam on the surface. This was usually brown, but sometimes red. As a rule the ants operated from a central hole around which they deposited a shallow pile of loose earth up to 6 in. across and 1 in. high, often so placed as to cover a leaf. The ants carry the soil particles in their mandibles, but do not attach each to its neighbour as the termites do. The mechanical composition of the brown earth deposited by a small black ant, and the red earth by a large black ant, are shown in Table 3. The proportions of clay to silt and fine sand in both samples show them to have been derived from the top 12 in. of soil. Although the superficial earth-moving activities of ants are overshadowed in this area by the earthworms, in other areas, where there are fewer worms, the topmost layers of the soil are formed largely of earth deposited by ants.

Earthworms

The importance of worms in forming a surface horizon of fine sandy loam has already been mentioned. Measurements were therefore made to determine the amount of material they deposited on the surface, the maximum depth from which it came, and their influence on the mechanical and chemical composition of the soil.

Our knowledge of the influence of worms on the soil has been recently reviewed (Anon., 1950). The amount of material deposited in casts on the surface was originally estimated by Darwin, who found that 7·6 and 16·1 tons per acre per year were deposited on two small areas of old turf. Evans (1948) has listed the rates of casting determined by several later workers, which are variable, but fall in general between his own limits of 25 tons per acre per annum on old pasture and 1 ton per acre per annum

on arable. The chemical composition of worm casts has been studied by Lunt and Jacobson (1944). They found the casts had a higher pH, more exchangeable cations, and more organic carbon than the topsoil. All these effects were attributed to the intimate mixing of plant and animal remains with mineral soil in the digestive tract of the worm, and to the

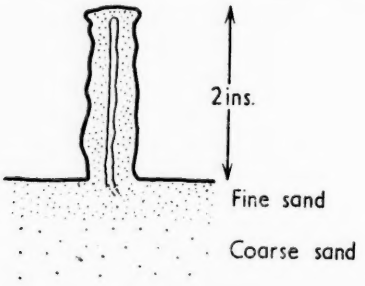


FIG. 2. Cast of *Hippopera nigeriae*.

action of digestive secretions on the mixture. In forest sites they found the ratio of exchangeable Ca to exchangeable Mg increased from 5·7 in the topsoil to 9·6 in the casts, but in a cultivated field the ratio decreased from 12·1 to 5·8. It seems likely that these differences can be attributed to differences in the composition of the organic matter on which the worms are feeding, but no analyses were given.

Description of the worms and their casts. The only species of earthworm that has been recorded in this area is *Hippopera nigeriae* (Taylor, 1949). It is similar to *Lumbricus* in external appearance and occurs in all sizes up to or exceeding 25 cm. long and 8 mm. in greatest diameter. It cannot be stated with certainty that all the smaller worms seen are members of this species, but only one type of cast occurs on the surface, of a most interesting form, and quite unlike that of the common British species of *Allolobophora*. The casts are irregular cylinders, up to 6 cm. long and 1·5 cm. in diameter, standing vertically on the surface of the soil and dark brown in colour. There is a small hole running through the centre of the cast almost to the top, which is sealed (see Fig. 2). The worm appears to lie just beneath the surface, and builds up the cast by extruding earth circularly around the tip of its tail, which it gradually pushes upwards through the extruded mass.

During the rainy season these worms are extremely active in the Ibadan area, but during the dry season there is no sign of casting except near streams. Their distribution is, however, very erratic, and there are many parts of the Gold Coast and Nigerian forest where there is little sign of worm activity. A photograph of the close carpet of casts left on the surface at the end of the rainy season has been shown in Part I. Over the catena studied, the earth from the decomposing casts was sufficient to form a well-defined layer ½–1 in. thick of fine sandy loam above the coarse sandy loam below, under both secondary forest and abandoned farmland covered with tall grasses.

A colourful description of the activities of these worms was given by Millson, Assistant Colonial Secretary of Lagos, in 1890. He writes:

> The whole surface of the ground among the grass is seen to be covered by serried ranks of cylindrical worm casts. These worm casts vary in height from a quarter of an inch to three inches, and exist in astonishing numbers. It is, in many places, impossible to press your finger upon the ground without touching one. For scores of square miles they crowd the land, closely packed, upright, and burnt by the sun into rigid rolls of hardened clay. There they stand until the rains break them down into a fine powder, rich in plant food, and lending itself easily to the hoe of the farmer. . . . On digging down, the soil is found to be drilled in all directions by countless multitudes of worm drills, while from 13 in. to 2 ft. depth the worms are found in great numbers in the moist subsoil.

By weighing casts on two separate square feet of land he estimated that 5 lb. per sq. ft. (equals 100 tons per acre) were brought up each year. His specimens were identified as 'a probably new species of the genus *Siphonogaster*'.*

Rate of casting. The casts formed each day on two plots, each of 25 sq. ft., were weighed for a month in the middle of the rainy season. The plots were under secondary bush about 25 years old, providing abundant leaf fall. Plot A was completely shaded, while plot B was about 75 per cent. shaded. In the month the dry weight of castings on plot A amounted to 4·05 tons per acre, and on plot B to 2·75 tons per acre. The soil remained moist throughout the period, and it is reasonable to assume that casting occurs at about this rate for the 6 rainy months of the year and ceases for the rest of the year. The average weight of casts formed during the year may therefore be estimated as 20·4 tons per acre, a figure that agrees well with the estimates of temperate areas.

The weight of worm-cast material and fine sandy loam topsoil derived from casts is about 120 tons per acre; so that, on the average, earth brought up by worms will be reincorporated in the coarse sandy layer below in about 6 years. This mixing will be due largely to the soil fauna.

The mechanical composition of the casts. The mechanical composition of the casts collected above the deep upper-slope profile, described in Part II, is compared with that of the upper layers of the profile in Table 3. It will be seen that the casts contain virtually no grains greater than 0·5 mm. and a low proportion between 0·2 and 0·5 mm. The maximum depth from which the earth in the casts is being obtained may be deduced from the ratio of clay to silt and fine sand in the casts; this corresponds to the ratio in the top 12 in. of the soil but not below. Probably the worms operate mainly in the surface inches.

The chemical composition of the casts. The organic-carbon content, pH, and base status of the casts are compared with those of the upper

* Millson concludes, amusingly in the light of our present knowledge: 'Of the effect of this constant moving of the soil upon the health of the country it is not possible for anybody but an expert to speak, but it seems more than probable that the comparative freedom of this part of West Africa from dangerous malarial fevers is due, in part at least, to the work of earth-worms in ventilating and constantly bringing to the surface the soil in which the malarial germs live and breed. It is, perhaps a temptation to attribute to them too many virtues, but it seems not unlikely that the beneficial effect of their labour in this direction would be very considerable.'

horizons of the soil below, in Table 4. The composition of the leaves collected from above the casts is also shown. The casts contain considerably more organic matter than the topsoil, and have a greater exchange capacity, saturation percentage, and pH. The exchangeable calcium, magnesium, and manganese contents are greater in the casts, but the exchangeable sodium and potassium are about the same. The proportions of exchangeable cations in the casts agree well with the relative amounts in the leaves overlying them. All these observations agree with the view that the casts are an intimate mixture of digested earth and plant remains.

TABLE 4

pH, Exchangeable Cations, Exchangeable Acidity, and Organic Carbon in Worm-casts compared with Soil below

Sample	pH	Exchangeable cations and acidity (m.e. per 100 g.)						Sat. %	Org. C. %
		Ca	Mg	K	Na	Mn	Acidity		
Worm-casts .	6·9	7·0	2·5	0·34	0·38	0·11	2·4	81	2·96
Cr W 0–1 in. .	6·2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0·96
Cr T 1–6 in. .	6·4	2·5	1·3	0·43	0·45	..	2·0	70	0·53
Cr G_1 6–12 in..	6·7	1·9	0·8	0·34	0·50	..	1·6	69	0·32
		total cations (m.e. per 100 g.)							
Leaves . .	..	84	26	5·3	2·6	0·77	..	..	..

Summary

The soil-forming activities of the soil fauna in the forest region of south-west Nigeria have been studied. Termites and ants have the most profound influence, and among them *Macrotermes nigeriensis* Sjost is the most important. Its mounds are described, and their mechanical and chemical compositions are compared with the surrounding soil. The maximum depth from which material has been brought up, the rate of formation and decay of the mounds, and their density of distribution are estimated. From these figures it is estimated that at least ½ ton of earth is deposited on the surface over an acre each year. It is further estimated that the minimum rate of lowering of the land surface is 1 ft. in 12,000 years.

The activities of the earthworm *Hippopera nigeriae* Taylor were also described. It is estimated that it deposits 20 tons per acre of earth in a year on the surface under forest in the form of cylindrical-shaped casts. The casts are richer in organic matter and exchangeable Ca and Mg, and have a higher pH, exchange capacity, and percentage saturation than the top 6 in. of soil.

REFERENCES

ADAMSON, A. M. 1943. Termites and the fertility of soils. Trop. Agric. Trin. **20,** 107.

Anon. 1950. The role of earthworms and insects in soil formation. Soils and Fert. **13,** 157.

ASHLEY, G. H. 1935. Studies in Appalachian Mountain sculpture. Bull. Geol. Soc. Amer. **46,** 1395–1436.

CLARKE, F. W. 1924. The data of geochemistry. U.S. Geol. Survey Bull. 770.

DARWIN, C. 1881. Earthworms and Vegetable Mould. Murray, London.
EVANS, A. C. 1948. Studies on the relationship between earthworms and soil fertility. Pt. II, Some effects of earthworms on soil structure. Ann. Appl. Biol. **35,** 1–13.
GRIFFITH, G. Ap. 1938. A note on termite hills. E. Afr. Agric. J. **4,** 70–71.
HARRIS, W. V. 1949. Some aspects of the termite problem. E. Afr. Agric. J. **14,** 151–5.
HEGH, E. 1922. Les Termites. Bruxelles.
KALSHOVEN. 1941. Influence of the local macroscopic fauna, especially the termites, on soil fertility (in Dutch). Tectona, **34,** 568–82. Quoted by Adamson (*v. supra*).
LUNT, H. A., and JACOBSON, H. G. M. The chemical composition of earthworm casts. Soil Sci. **58,** 367–75.
MARAIS, E. N. 1950. The Soul of the White Ant. 6th edition, Methuen, London.
MILLSON, A. 1890. Soil and cultivation in Yoruba-land. Kew Bull. No. 37, 238–44.
MILNE, G. 1947. A soil reconnaissance journey through parts of Tanganyika Territory, December, 1935 to February, 1936. J. Ecol. **35,** 192–265.
PENDLETON, R. L. 1942. Importance of termites in modifying certain Siamese soils. J. Amer. Soc. Agron. **34,** 340–4.
SHALER, N. S. 1891. The origin and nature of soils. U.S. Geol. Survey 12th Ann. Rep. 1890–1 (pt. i), 213–345.
TAYLOR, A. G. 1949. A West African earthworm. Proc. Zool. Soc. Lond. **119,** 703–10.

THE EFFECT OF SOIL-FORMING FACTORS OVER AN AREA IN THE SOUTH OF SCOTLAND

JOHN W. MUIR

(*The Macaulay Institute for Soil Research, Aberdeen*)

Introduction

THE area to be discussed was surveyed by the Soil Survey of Scotland during the summer months of 1949–53. The principal surveyors were J. W. Muir (1949–52), M. J. Mulcahy (1952), and J. M. Ragg (1953). In the memoir of the area which is at present being prepared, the soils are mapped as Soil Series which are placed into Associations. For this paper, however, the Soil Series are considered as members of major soil groups. The principal major soil groups of the area are Brown Forest Soils of low base status, Non-calcareous Gleys, Peaty Podzols with iron pan, and Peaty Gleys. Hill Peat is also very common.

The purpose of this paper is to attempt to explain the occurrence and distribution of the major soil groups. The five soil-forming factors, climate, parent material, topography, vegetation, and time are discussed and an assessment made of their influence.

General Description of the Area

The area surveyed amounts to approximately 500 sq. miles. It consists of most of Roxburghshire and small parts of Selkirkshire and Dumfriesshire. The principal river is the Teviot which, with its tributaries, drains practically the whole area; only the Bowmont Water in the east, the Liddel Water in the south, and the Ettrick Water in the north-west are not tributaries.

There are three principal types of terrain: the mountains and high-level plateaux of the southern and eastern parts of the area, the upland moors of the west, and the hilly to rolling country of the north and centre. The physical map (Fig. 1) shows this clearly.

Climate

Fig. 2 is a rainfall map of the area. The main trends are (i) increasing rainfall towards the source of the prevailing south-west wind which brings by far the greatest part of the rainfall, and (ii) increasing rainfall with increasing altitude.

Only a few facts about the temperature and humidity of Hawick are given. The average mean of daily mean temperature (1906–35) for January is 37° and for July is 57° F. The average mean of daily range of temperature (1906–35) for January is 10° and for July is 17° F. The average mean of relative humidity at 13H (1921–35) for January is 83 per cent. and for July is 68 per cent.

[Journal of Soil Science, Vol. 6, No. 1, 1955.

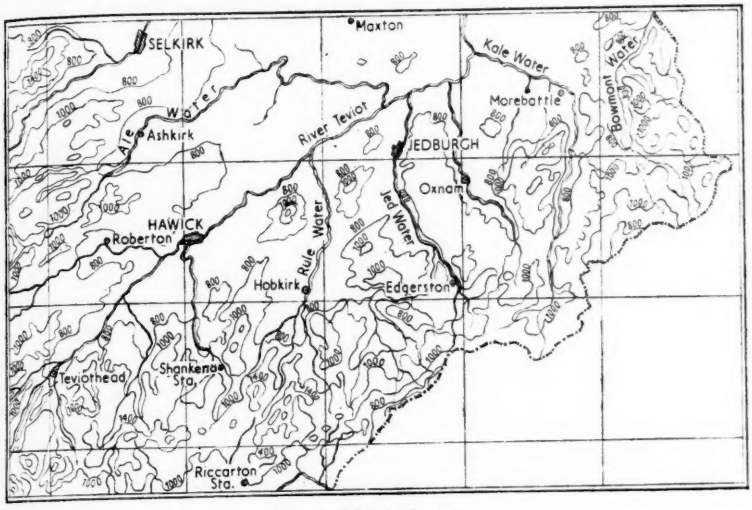


FIG. 1. Physical map.

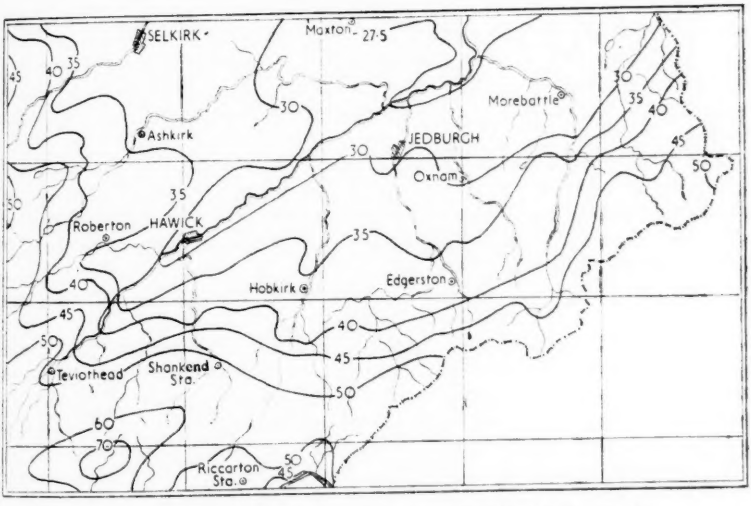


FIG. 2. Distribution of annual average rainfall (1881–1915).

Surface Geology

The whole of the area has been glaciated and, in general, is covered with tills. They are relatively local in origin and their distribution is very similar to that of the solid formations from which they have been derived. There has been very little mixing of materials from two or more solid formations: only in the neighbourhood of the junction of the

Wenlock and Upper Old Red Sandstone sediments on the one hand and Llandovery and Upper Old Red Sandstone sediments on the other has it been necessary to separate a till of mixed geological origin.

Fig. 3 shows the distribution of the various tills. The character of the parent rock is reflected in that of the till. That derived from Lower Old Red Sandstone lava occurs in the eastern part of the area. The lavas are intermediate in composition except for a small area of rhyolite which occurs on the watershed between Cocklaw and Rushy Fell. Till derived from Upper Old Red Sandstone Sediments occurs in the central part of the area. The sediments are principally red sandstones and mudstones, but they vary somewhat, and white siliceous strata become more common towards the top of the formation. Till derived from Lower Carboniferous Sediments covers two parts of the area near the watershed. The sediments comprise cementstones, mudstones, shales and, towards the top of the formation, siliceous sandstones. Till from Llandovery rocks occurs in the western part of the area. The rocks consist of greywackes, mudstones, and shales. They are very highly folded, and in certain localities where the till is thin a series of ridges known as corrugations is formed. They have invariably a south-west to north-east strike. Rock, especially greywacke, tends to outcrop at the apex of each corrugation. Till derived from Wenlock rocks occurs in the south-western part of the area and in certain isolated areas farther east. Like the Llandovery, the rocks consist of greywackes, mudstones, and shales with possibly less greywacke. The folding does not seem to be so marked and certainly there are no areas of corrugations. Fluvio-glacial sand and gravel are to be found along the lower courses of the Teviot, Kale, and Ettrick Waters.

One very noticeable feature is that there are two tills formed from each parent rock. One is lighter-textured and more stony than the other. The lighter-textured till is normally found on the sides and shoulders of hills, and on high-level plateaux; the heavier-textured till is found in valleys and on smooth slopes.

Soils

During the survey, the soils were mapped as Soil Series, each series having a characteristic profile. Since in this paper the soils are discussed from the genetical point of view, it was considered desirable to place the Soil Series into higher categories, the major soil groups. Four major soil groups occur in the area; Brown Forest Soils of low base status, Peaty Podzols with iron pan, Non-calcareous Gleys, and Peaty Gleys. In addition there is Hill Peat. These major soil groups correspond to certain mentioned by Muir and Fraser (1940). Brown Forest Soils of low base status are the equivalents of their Brown Earths, Peaty Podzols with iron pan the equivalents of Peat Podzolized Soils, Non-calcareous Gleys the equivalents of Surface Water Gleys, and Peaty Gleys the equivalents of Peaty Gleys. It has been less easy to identify all the major soil groups in Kubiena (1953). Brown Forest Soils of low base status correspond to his Cryptopodzolic Braunerde and Non-calcareous Gleys to his Pseudogleys, but the other two do not seem to fit any of the other

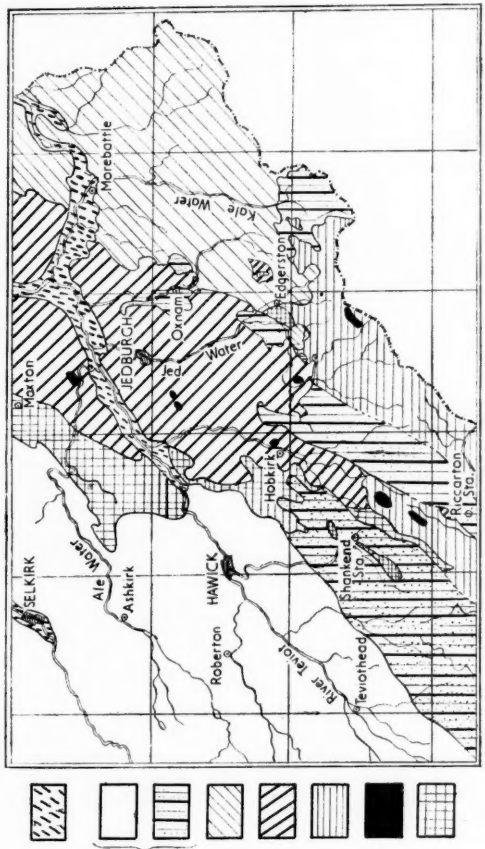


FIG. 3. Generalized distribution of glacial tills.

specifications exactly. Peaty Podzol with iron pan corresponds most closely to Peat Podzol and Peaty Gley to Molken Podzol.

Profile descriptions of the major soil groups occurring on till derived from rocks of the Llandovery Division of the Silurian are given. They are typical examples of the major soil groups found over the whole area. All the Soil Series mapped on tills derived from the other rocks have their own series characteristics but, nevertheless, they belong to the same major soil groups.

Profile descriptions:

Major soil group: Brown Forest Soil (low base status).

Topography . . . Short, sharp slope.
Aspect . . . . South.
Altitude . . . 1,000 ft.
Vegetation . . . *Agrostis* spp., *Festuca* spp., *Pteridium aquilinum*, Hypnaceous mosses.
Drainage . . . Free.

L ½ in. Grass and moss litter.
F ½ in. Partly decomposed litter.
H 1 in. *Dark brown humus; live roots penetrating.
A 0–10 in. Yellowish-brown silty loam; fine cloddy; low organic matter; stone content moderate; dry; no mottling. Merging into
B_2 10–14 in. Yellowish-brown silty loam; fine cloddy; low organic matter; moderate stone content; dry; no mottling; clear change into
B_3 14–17 in. Light olive-brown, silty clay loam; soft cloddy; no organic matter; stone content moderate; dry; no mottling; sharp change into
C 18+ in. Light olive-brown, silty clay loam; very many stones with finer material in interstices; no organic matter; few roots; moist. No mottling.

Major soil group: Peaty Podzol with iron pan.

Topography . . . Moderately gentle slope.
Aspect . . . . North.
Altitude . . . 1,100 ft.
Vegetation . . . *Molinia caerulea*, *Calluna vulgaris* with *Polytrichum commune* and *Juncus squarrosus*.
Drainage . . . Poor above iron pan; free below iron pan.

L ½ in. Grass litter.
F ½ in. Partially decomposed grass litter.
H 6 in. Black mealy humus. Live roots penetrate.
A_1 0–3 in. Dark grey-brown, silty loam with dark brown patches of organic staining; cloddy structure; stone content moderate; roots matted above the iron pan; moist; no mottling. Sharp change into
B_1 at 3 in. Thin iron pan, well defined and continuous.
B_2 3–10 in. Pale brown silty clay loam; cloddy structure; no organic matter; stones abundant; roots rare; moist; strong ochreous brown and grey mottling, rare pinheads of black staining; merging into
B_3–C 10+ in. Reddish-brown, gritty silty clay loam; very compact; stones abundant with increase in red shale content; moist; brown and ochreous mottling; more black staining than above; mottling and staining decrease with depth.

Major soil group: Non-calcareous Gley.

Topography . . . Gentle slope.
Aspect . . . . North.
Altitude . . . 950 ft.
Vegetation . . . *Calluna vulgaris*, *Nardus stricta*, with *Sphagnum* and other mosses.
Drainage . . . Poor.

* Munsell colour names are used in the descriptions.

L F		Trace.
H	½ in.	Very dark brown humus.
A_1	0–5 in.	Dark grey-brown clay loam; cloddy; low organic matter; few small stones; wet; slight ochreous mottling round roots. Sharp change into
B_1	5–8 in.	Light olive-brown, silty clay; cloddy to prismatic; no organic matter; few small stones; moist; strong bright orange mottling; sharp change into
B_2	8–16 in.	Light olive-brown silty clay; cloddy; no organic matter; numerous stones; moist; light grey-brown coloured material round stones and roots; ochreous mottling; merging into
B_3–C	16+ in.	Light olive-brown silty clay; cloddy; no organic matter; many stones; dry; grey mottling round stones; grey channels to roots; grey mottling decreases with depth; some stones soft and ochreous coloured.

Major soil group: Peaty Gley.

Topography . . .	Very gentle slope.
Aspect . . .	North.
Altitude . . .	1,000 ft.
Vegetation . . .	*Molinia caerulea, Trichophorum caespitosum, Calluna vulgaris.*
Natural drainage .	Poor.

L	½ in.	Undecomposed plant litter.
F	1 in.	Partially decomposed plant litter.
H	4½ in.	Black humus.
A_1	0–4 in.	Olive-grey silt loam; cloddy to prismatic structure; uniform brown organic staining, decreasing towards base of horizon; stones rare; moist; strong brown mottling in small patches and around root channels; merges into
B_1	4–9 in.	Light olive-grey silt loam; cloddy to prismatic structure; stones rare; mottling as above but slightly stronger; sharp change into
B_2	9–20 in.	Light brown-grey silty clay loam; large cloddy structure; stones occasional, some rotten; moist; strong orange and brown mottling; grey root channels and faces to stones and structural units; merging into
B_3–C	20+ in.	Light brown-grey silty clay; large cloddy structure; stones frequent; moist; mottling as previous horizon but stronger neutral grey colours.

Hill Peat is also an important soil in this area. A soil is considered Peat when it has a depth of raw humus of more than 15 in. Very often there is either iron pan in the mineral soil under the peat or the peat rests on solid rock.

Distribution of Major Soil Groups

Fig. 4 shows the generalized distribution of the major soil groups and Hill Peat. In general, Hill Peat, Peaty Podzols, and Peaty Gleys are to be found near the watershed and on the western uplands. Brown Forest Soils and Non-calcareous Gleys are present in the central part of the area.

Two minor effects, which it has been impossible to show, due to considerations of scale, deserve mention. Brown Forest Soils do occur in areas shown in Fig. 4 as Peaty Podzols and Peaty Gleys, but this is an effect of topography and will be discussed under that heading. Likewise Non-calcareous Gleys occur within the same areas, but only to a very

small extent. No explanation of this can be given except to say that some but not all of them are 'flush' types.

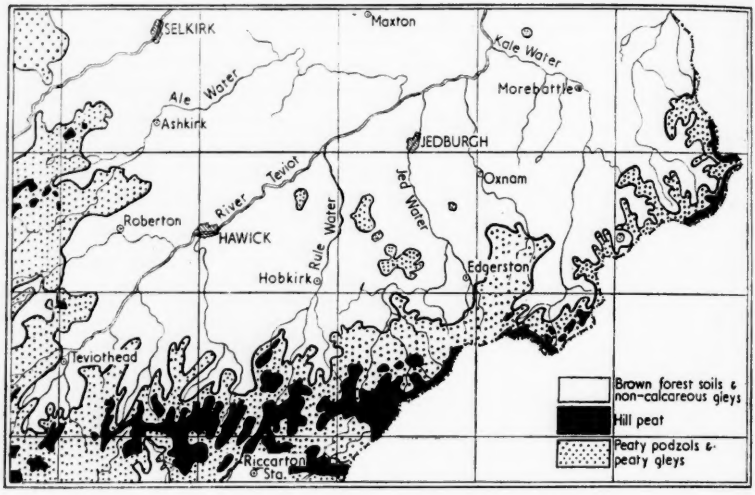


FIG. 4. Generalized distribution of major soil groups.

Discussion

The factors (Jenny, 1940) affecting the distribution of major soil groups are (1) climate, (2) parent material, (3) topography, (4) vegetation, and (5) time. These will be discussed in turn and an assessment made of their effect on the major soil groups of this area.

Climate plays a decisive part in the occurrence and distribution of the four major soil groups. Rainfall is the most important climatic element in this area and it should be noted, at the outset, that the rainfall gradient towards the watershed is very steep. Fig. 4 shows the areas where Peat, Peaty Podzols, and Peaty Gleys, on the one hand, and Brown Forest Soils and Non-calcareous Gleys on the other are predominant. By comparing Figs. 2 and 4 it can be stated with some certainty that Peat, Peaty Podzols, and Peaty Gleys generally occur under conditions of high rainfall. When Brown Forest Soils occur under high rainfall, as they do to some extent, they are to be found only on steep slopes: this will be discussed later under Topography. In addition, as can be seen from Fig. 4, there is a tendency for Peaty Podzols and Peaty Gleys to occur under low rainfall in the area of till derived from Upper Old Red Sandstone sediments, but this is a parent-material effect and will be discussed later. These minor effects aside, it is true to say that with increase in rainfall there is a tendency (1) for raw humus to accumulate and (ii) for iron pan to be formed.

Parent material plays an important part in the distribution of the major soil groups, both by virtue of its texture and by virtue of its base-richness. The effect of texture will be discussed first.

The texture of the parent material in this area has determined to a large extent the natural drainage of the soils. If the soil is developed on a heavy-textured till it is a gley, either a Peaty Gley or a Non-calcareous Gley, while if it is developed on a light-textured parent material it is either a Peaty Podzol or a Brown Forest Soil. This effect is most clearly seen near the boundary of the tills derived from Llandovery and Upper Old Red Sandstone rocks. On the Llandovery till which is predominantly heavy-textured in this locality, 90 per cent. of the soils are Non-calcareous Gleys, while on the till derived from Upper Old Red Sandstone sediments which is predominantly light-textured, the soils are 90 per cent. Brown Forest Soils. The texture of a till depends on the nature of the parent rock and the effects of glaciation. Two tills, one lighter-textured and more stony than the other, have been formed from all the parent rocks. The lighter-textured till occurs on the higher slopes and high-level plateaux while the heavier-textured till occurs on the lower slopes and in valleys. This separation into two tills can be attributed to glaciation because it occurs on all the parent rocks. It should be noted, however, that the proportion of lighter-textured till differs in each case and therefore this can be considered an effect of the parent rock. The Llandovery rocks have yielded for the most part a heavy-textured till, the lighter-textured till occurring only where there is an area of corrugations. The Wenlock and Lower Carboniferous sediments have yielded about equal amounts of light-textured and heavy-textured till. The Lower Old Red Sandstone lavas have yielded two tills, but the light-textured is much more common than the heavy-textured. The Upper Old Red Sandstone sediments have yielded almost all light-textured till with very little heavy-textured till to be found. Another characteristic of the Upper Old Red Sandstone till, not shared by the others, is its shallowness throughout the area: the lower part of a profile is often rotten sandstone. To sum up, glaciation has produced in this area two tills, differing in stoniness and texture, from each parent rock, but the type of parent rock has determined, in each case, the ratio of light-textured to heavier-textured. The soils on heavy-textured tills are Gleys, while those on light-textured tills are either Brown Forest Soils or Peaty Podzols. This is one of the most important effects of parent material.

The base-richness of the parent rock is also important, but its effect is less clear-cut in this area than elsewhere. It plays a part in determining the distribution of Peaty Podzols and Peaty Gleys on the one hand and Brown Forest Soils and Non-calcareous Gleys on the other. The tills derived from all the parent rocks except parts of the Upper Old Red Sandstone and Lower Carboniferous sediments are unsuitable for study because of their intermediate composition. The Lower Carboniferous sediments occur only under high rainfall and this masks the parent-material effect. The Upper Old Red Sandstone sediments, however, vary in base-richness, the strata towards the top of the formation tending to become more siliceous. Where this happens in the area of low rainfall Peaty Podzols and Peaty Gleys tend to occur. Nearby under otherwise similar conditions, but the till being less siliceous, there are Brown Forest Soils and Non-calcareous Gleys. Fig 4 shows this clearly, small

isolated areas of Peaty Podzols and Peaty Gleys occurring under low rainfall on till derived from Upper Old Red Sandstone sediments.

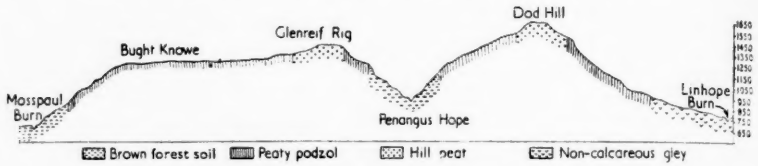


FIG. 5. Vertical section through part of the area showing major soil groups.

Topography plays an important part in the distribution of Hill Peat, Peaty Podzols, and Brown Forest Soils in areas of high rainfall. Peaty Podzols and Brown Forest Soils invariably occur on light-textured till while Hill Peat sometimes occurs on light-textured till with an iron pan or sometimes on solid rock. In such conditions of high rainfall Hill Peat occupies the flatter tops, Peaty Podzols the moderate slopes, and Brown Forest Soils the steep slopes: the steepness of slope occupied by each type varies with the rainfall. Fig. 5 shows this clearly. It is a section through part of the area covered by till derived from Wenlock rocks. Hill Peat is on the flatter tops, giving way to Peaty Podzols on moderate slopes and finally Brown Forest Soils on steep slopes. Finally on a moderate slope where Peaty Podzols are predominant Brown Forest Soils will nevertheless occur wherever there is a micro-slope sufficiently steep.

Robinson (1935) and Jenny (1941) do not consider the vegetation of a soil to be an independent variable. Experience in this area bears this out. The most that can be said is that a certain type of vegetation can often be associated with each major soil group (Stremme, 1926; Muir and Fraser, 1940). Brown Forest Soils have usually some form of acid grassland with *Agrostis* spp., *Festuca* spp., and *Pteridium aquilinum* prominent. Peaty Podzols have usually some form of Molinetum, either *Molinia caerulea—Calluna vulgaris* or *Molinia caerulea—Nardus stricta*. Non-calcareous Gleys have not always the same type of vegetation. Sometimes *Deschampsia caespitosa* is prominent and at other times the vegetation is *Calluna vulgaris* with *Nardus stricta* and *Sphagnum*. Peaty Gleys have invariably some form of Molinetum. Hill Peat has usually either *Eriophorum* spp. dominant or, where it is eroding, *Calluna vulgaris* with often *Rubus chamaemorus*.

Finally, time as measured by years or centuries does not play a part in differentiating these soils, for they were all freed from glaciation at approximately the same time.

Conclusions

Climate has had an important influence on the character of the soils in this area. Generally speaking, Hill Peat, Peaty Podzols, and Peaty Gleys occur where the rainfall is high and Brown Forest Soils and Non-calcareous Gleys where the rainfall is lower. Brown Forest Soils certainly occur under high rainfall, but only where the topography is suitable.

Parent material has two noteworthy influences. Its texture largely determines whether a soil is a gley or not and its base-richness affects the distribution of Peaty Podzols and Peaty Gleys. A type of vegetation characteristic of each major soil group has been distinguished, but it is doubtful if vegetation has had a primary effect on the type of soil. Time has not had any noticeable effect.

Summary

A general description of the area is given with data on the climate and geology. The major soil groups to be found are described, and the influence of each soil-forming factor on their occurrence and distribution is discussed.

REFERENCES

JENNY, H. 1941. Factors of Soil Formation. McGraw-Hill Book Company Inc.

KUBIENA, W. L. 1953. The Soils of Europe. Thomas Murby & Co.

MUIR, A., and FRASER, G. K. 1940. The soils and vegetation of the Bin and Clashindarroch forests. Trans. Roy. Soc. Edin. **60,** part i, 233–341.

ROBINSON, G. W. 1935. Soils of Great Britain. Trans. Third Intern. Congr. Soil Sci. **2,** 11–23.

STREMME, H. 1926. Grundzüge der praktischen Bodenkunde. Verlag von Gebrüder Bomtraeger, p. 83.

A REVIEW OF THE MINERALOGY OF SCOTTISH SOIL CLAYS

W. A. MITCHELL

(*The Macaulay Institute for Soil Research, Aberdeen*)

In the course of routine examination of Scottish soil clays by X-ray diffraction a large variety of soils has been examined and an attempt can now be made to correlate clay mineralogy with parent material. There are many difficulties involved in such an undertaking. With the exception of a few special cases the clay minerals present in soils are few in number and widespread in occurrence. The differences that are observed are very often only in the relative amounts of the minerals present, so that correlation with the type of parent material is dependent not only on qualitative identification but to a large extent also on quantitative estimations.

In Table 1 the composition of the clay in the surface soil of profiles developed on various parent materials is given. The figures, which indicate the approximate ratios in which the clay minerals are present, are the averages of all the profiles examined, the actual number of profiles concerned being shown in the third column. No allowance has been made for amorphous material, for oxides or hydroxides of iron and aluminium, or for any other minerals present. Drainage classes, as defined by the Scottish Soil Survey, are indicated by the symbols in the second column in which P represents the freely drained, PH the moderately gleyed, and HP the strongly gleyed profiles. The direction of the arrows after the figures indicates a tendency for the mineral content to increase towards the top or bottom of the profile.

TABLE 1

Parent material	*Drainage*	*No of profiles*	*Illite*	*Kaolin*	*Montmorillonite*	*Vermiculite*	*Chlorite*
Olivine Basalt	P	2	2	1	4	3	0
Norite	P	4	1↓	4	0	5↑	0
Norite	PH	3	1	3	4↓	2↑	0
Diabase	P	1	2	2	1	5	0
Diabase	PH	1	2	2	5	1	0
Andesite	P	1	5↓	3↑	0	2↓	0
Andesite	HP	1	3↑	2↑	3↓	2↑	0
Granite	P	10	7↓	3↑	0	0	0
Granite	PH	8	6↓	3↑	0	1↑	0
Quartz-Schist	P	1	8	2	0	0	0
Quartz-Schist	PH	2	6↓	3↑	0	1↑	0
Quartzite	P	1	5	3	0	2	0
Sandstone (Old Red)	P	4	6↓	3↑	0	1	0
Sandstone (Old Red)	PH	4	6↓	3↑	0	1	0
Slate (Dalradian)	P	2	5↑	3↑	0	2↓	0
Slate (Dalradian)	PH	2	6	2	0	2	0
Shale (Silurian)	HP	2	5↓	2↑	0	0	3↑
Clay (Jurassic)	PH	1	7↓	3↑	0	0	0
Clay (Interglacial)	HP	2	5↓	2	3↑	0	0

The heading of the first column in Table 1 might more accurately be *Grandparent material*, as most Scottish soils are developed from drift

deposits derived from solid-rock formations. Care was taken in the selection of the profiles quoted to exclude soils derived from material which was obviously of mixed origin, but some contamination is probably present in most cases.

The identification of the clay minerals by X-ray diffraction is based on the first-order basal or c-spacing and on the (06) line or sixth order of the b-spacing, and on the effects on these lines of glycerol and heat treatments. These are standard techniques and will not be described in detail here.

The illites and montmorillonites can occur either as dioctahedral or trioctahedral types. These are analogous to the rock minerals muscovite and biotite respectively. The distinction between those two types is easy in a pure mineral, but it may be very difficult in the case of a soil clay containing both, owing to the overlapping of the (06) lines. No distinction is made between the two types of illite and montmorillonite in Table 1. In the case of illite the dioctahedral type is by far the more common, but the trioctahedral type has been identified in some of the soils on basic igneous rocks.

The identification of the kaolin mineral in soil clays is usually only possible if it is present in appreciable amounts. Where identification is possible the mineral is very often intermediate between kaolinite and halloysite and similar to 'fireclay' (Brindley and Robinson, 1947). Partly hydrated halloysite has been found in soils on basic igneous rocks, and the kaolin mineral in the quartzite soil profile is well crystallized, approaching a true kaolinite. There seems to be a correlation between the type of kaolin mineral in the soil and the acidity of the parent rock, increasing acidity giving increased order in the kaolin crystals.

Most of the quantitative results quoted in Table 1 were arrived at by a visual comparison of the relative densities of the lines on photographs of the soil clays with those on photographs of standard mixtures of clay minerals in various proportions. This method has the disadvantage that it only gives relative amounts. An improved method which is now in use is based on the addition of potassium chloride to the soil clays as an internal standard. The relative intensities of the soil clay-mineral and potassium-chloride lines are compared with those of standard mixtures of pure clay minerals and potassium chloride. The results obtained for each mineral by this method are absolute, and are independent of the other material present whether crystalline or amorphous. This method was used in only a few of the profile analyses in Table 1.

Both these methods of quantitative analysis have one defect. They are based on the assumption that the standard minerals give the same diffraction pattern as the corresponding minerals in soil clays. The considerable range in chemical composition found in montmorillonites and illites gives rise to differences in the relative intensities of their diffraction lines (Brown, 1951). Considerable errors may arise in this way.

Table 1 has been arranged with the igneous rocks in order of increasing acidity at the top. In this way the soils can be divided at the level of the andesite profiles into two groups which show marked differences. Above the division are the rocks containing large amounts of ferromag-

nesian minerals. These give rise to brown earths and associated gley soils. Their clay fraction is dominated by vermiculite or montmorillonite, with a tendency for the former to prevail under free drainage and the latter in the gleys. This influence of drainage is clearly seen in soils developed on norite, where the change from vermiculite to montmorillonite can be traced on going up the moderately gleyed profiles. The lower layers contain montmorillonite often associated with trioctahedral illite, and the amount of these minerals becomes progressively less on going upwards into the more freely drained layers, with a corresponding increase in vermiculite.

The predominance of illite in the lower half of the table is striking, as is also its tendency to increase towards the foot of the profiles. The content of kaolin is usually about half that of illite and often shows an increase towards the top of the profile. The influence of drainage differences on the clay mineralogy of soils on granitic and sedimentary rocks is less evident that on basic igneous rocks. There is, however, a tendency for the increase in the kaolin content towards the surface to be more pronounced in the freely drained soils than in the gleys. Scottish soils on granite are all strongly influenced by one or both of two opposing factors, podzolization and gleying. The diffuse B_2 horizon of the classical podzol occurs under coniferous woods, while under wet heath vegetation thin iron pans are formed which may be so well developed that they impede the drainage, giving gleying conditions in the surface. Surface gleying is also caused by the water-holding capacity of a thick layer of mor humus. Gleyed profiles also result from high local or regional water tables. Although a large variety of profiles may be found under these conditions the clay-mineral content of the soils remains fairly constant with about two or three parts of illite to one of kaolin.

The only soils examined which contain appreciable amounts of chlorite in the clay fraction are those on Silurian shale which contains macroscopic chlorite. A grain-size separation of the clay from one of these profiles showed the proportion of chlorite in the finest fraction, i.e. $0{\cdot}13\,\mu$, to be the same as in the total clay. This fact and a marked increase in the chlorite proportion in the upper layers suggest a high degree of stability for chlorite in soils. This is the only mineral in these soils that occurs both as large single crystals in the parent material and in the fine clay, and is a major constituent in both cases. The montmorillonite in the soil on Interglacial clay is also derived directly from the parent material.

The mechanism of soil formation is very complex and great care must be taken in attempting to interpret the trends seen in the changes in the clay-mineral contents of the various horizons of a soil profile. The increase we find in the amount of any mineral from one horizon to another is only relative to the other minerals present, and does not necessarily represent an absolute increase; there may in fact be an absolute decrease, the mineral in question being decomposed less quickly than the other components of the clay. Changes in the relative amounts of two minerals in a soil profile may arise in a number of ways. It may be that both minerals are being formed or destroyed at different rates; the absolute content of

either may be constant while the other changes; or one of the minerals may be changing into the other.

Absolute changes in the amounts of soil minerals could only be estimated if a suitable yardstick were available for comparison, such as a very stable mineral uniformly distributed through the parent material. It might be possible to use the magnetite content of some igneous rocks in this way, and by comparing its content in the parent material with that in the soil horizons to get some idea of the amount and nature of the material which is lost in solution and suspension during soil formation.

Soil minerals have in general different optimum conditions for formation and stability. Such variables as moisture content, drainage, aeration, temperature, pH, organic matter, &c., control the products of alteration of the parent material. These variables often change in a discontinuous way throughout a soil profile, so that a mineral which is stable in one horizon may be rapidly altered or decomposed in the next one.

An interesting feature which illustrates this point was found in the clay fraction from the freely drained podzol profile on Quartz Schist in Table 1. The relative amounts of illite and kaolin, as judged from the basal reflections, showed no variation throughout the profile; but the (06) reflections indicated dioctahedral and trioctahedral minerals in approximately equal proportions in the lower layers, with a marked decrease in the relative amount of the trioctahedral type above the B_2 layer. This suggests that in the acid and iron-free surface layers trioctahedral illite is unstable. No other podzol profile containing trioctahedral clay minerals has been found.

With regard to the alteration of individual rock minerals in pedogenesis much has still to be learnt. The examination of single crystals taken from weathered igneous rocks has given some useful information, and the changes found in this way are in general agreement with the results in Table 1. Among the ferromagnesian minerals it has been shown that biotite alters to vermiculite (Walker, 1949; Stephen, 1952), and hornblende to vermiculite and chlorite (Stephen, 1952). Unpublished work by the author has shown olivine crystals from weathered basalt to be altered to a mixture of trioctahedral montmorillonite and goethite. The weathering of muscovite to form illite is readily understood from the close similarities of their structures, but the formation of illite and kaolin from felspars (Milne, 1952) is a much more fundamental change involving the rearrangement of silicon-oxygen tetrahedra from a three-dimensional framework to the two-dimensional sheets of the clay minerals. An intermediate stage of something approaching true solution must occur with SiO_2 and Al_2O_3 existing as dispersed colloids which are precipitated to form amorphous gels from which kaolin crystallizes.

The clay minerals in soils can thus be grouped into three categories according to their origin. In the first group are those which are inherited directly from the parent material. These are most common in soils derived from sedimentary rocks which are often composed almost entirely of clay minerals. They may suffer little alteration during pedogenesis beyond a purely mechanical breakdown to a smaller particle size

as in the case of some chlorites. Secondly, there are the clay minerals which are formed by an alteration of rock minerals of similar structure, much of the original structure being retained as in the formation of illite from muscovite and vermiculite from biotite. The third group contains those minerals which are synthesized from the dissolved and amorphous products of weathering of the parent material; and here the final structures may have no relationship to those of the parent minerals.

Summary

A comparison of the results of quantitative mineralogical analyses of Scottish soil clays from 51 profiles shows significant differences between soils on basic igneous rocks on one hand, and on granite and various sedimentary and metamorphic rocks on the other. The former group is characterized by the predominance of montmorillonite or vermiculite, while in the latter illite predominates. The influence on clay mineralogy of drainage differences in soil profiles is more pronounced in soils on basic than on acidic parent materials. The weathering products of some common rock minerals are discussed with reference to the analytical results. Soil clay minerals are referred to three categories: those inherited from the parent material, those formed by alteration of the parent minerals, and those synthesized in the soil.

REFERENCES

BRINDLEY, G. W., and ROBINSON, K. 1947. Fireclay. Trans. Brit. Ceramic Soc. **46**, 49–62.

BROWN, G. 1951. X-ray Identification and Structures of Clay Minerals, pp. 155–72. The Mineralogical Society, London.

MILNE, A. A. 1952. An investigation of the mineralogical changes involved in the weathering of granites in the vicinity of Aberdeen. Thesis, University of Aberdeen.

STEPHEN, I. 1952. A study of rock weathering with reference to the soils of the Malvern Hills. Part i. J. Soil Sci. **3**, 20, and part ii, ibid. **3**, 219.

WALKER, G. F. 1949. The decomposition of biotite in the soil. Min. Mag. **28**, 693.

THE HYDRATION AND DEHYDRATION OF NA-MONTMORILLONITE (BELLE FOURCHE)

N. COLLIS-GEORGE

(*School of Agriculture, Cambridge*)

Introduction

THE aim of the experiments here reported was to make use of developed, or easily modified, experimental techniques, viz. tension plate, pressure membrane, and vapour pressure measurements, to produce a complete moisture-energy relationship for a single material, and with this information to consider the mechanism and kinetics to be attributed to wetting and drying in a swelling clay. Owing to the time limitation imposed on these experiments,* this latter consideration was not sufficiently resolved and will not be pursued in this paper. However, sufficient measurements were made to allow consideration of the methods of water removal and addition in swelling clays, at room temperature.

Belle Fourche Montmorillonite

The clay, provided in raw lump form by the American Colloid Company, is described by the American Petroleum Institute (1951) as Standard Mineral 27, Belle Fourche Montmorillonite, South Dakota. Its base-exchange capacity is 85–86 m.e. per cent., of which some 73 m.e. per cent. is exchangeable sodium and the remainder exchangeable calcium. On being shaken in distilled water, this clay is self-dispersing. All information available suggests that the clay of less than 1 micron in diameter, apart from a trace of gypsum, is entirely montmorillonite.

By sedimentation, particles of equivalent diameter greater than 1 micron were removed. The resultant suspension was divided into three parts: (i) Evaporated to dryness at 105° C., lightly ground to pass a 200-mesh sieve, and stored at 105° C. When required, it was cooled in a desiccator. (ii) First allowed to settle under gravity and then concentrated on the pump. The tension of the resultant paste was approximately zero, whilst its water content was 1,420 per cent. by weight, with a pH of 8·20 as measured by a glass electrode-calomel half-cell system. (iii) Coagulated with NaCl to give a resultant solution 0·45 N NaCl, concentrated by filtration to a paste, at zero tension, of pH 8·06 and of 1,600 per cent. water content.

Experimental Procedures

1. *Tension-plate apparatus* (Haines, 1930)

The sintered disk, which acts as a permeable membrane to aqueous fluids, is impermeable to air or water-insoluble fluids until a limiting suction pressure corresponding to the air-entry value of the largest pore is exceeded.

* The author carried out the experimental work, whilst on sabbatical leave, and wishes to acknowledge his indebtedness to the Department of Soils, University of California, Berkeley.

A closed system, with the vapour above the sample and disk at equilibrium with the vapour above the hanging column, which exerts the pressure deficiency on the water in the sintered disk, minimized evaporation losses. Flexible tubing connected the top of the Buchner sintered funnel to the manometer of the hanging column: the rest of the system, which was rigidly supported, was of pyrex. A metre vernier cathetometer, at 5 ft. distance, was used to determine applied suctions and volumes of water absorbed or released by the clay at each applied suction. The manometer of each apparatus was directly calibrated (1 cm. being approximately 0·1 c.c.).

By means of a three-way tap, which isolated the sample (and sintered disk), fluid could be added or subtracted from the manometer. Using a disk of nominal pore size 15 microns (air-entry value of tenth atmosphere), suctions up to 90 cm. water were obtained with the apparatus filled with water or 0·45 N NaCl. For a disk of nominal pore size 5 microns, suctions up to 450 cm. water could be obtained for both water and 0·45 N NaCl equilibriations by using a mercury hanging column.

The results, later discussed, were obtained in duplicate from a set of two water and two water-mercury apparatus. With the apparatus filled with the appropriate aqueous medium, dehydration curves were obtained in the conventional way from these clay pastes. Hydration curves were obtained by first draining the empty sintered funnel to suctions approaching the pore-entry value and then adding a 1 mm. layer of cooled oven-dry clay. Water (or salt water) was taken up by the clay and the resultant suction and aqueous adsorption observed by the cathetometer. By use of the tap, the clay was isolated whilst more liquid was added to the manometer. On reconnecting the sample and manometer the resultant lower tension and amount of water absorbed to produce this change were determined by cathetometer. The time for equilibrium varied with the suction, but 12–24 hours was generally observed to suffice.

2. *Pressure membrane apparatus* (Richards, 1947)

Pressure cells, whose construction depended on the use of rubber 'O' rings, were used in conjunction with Visking cellulose casing, 0·0045 in. thick, and were found to allow working pressures up to 150 atmospheres without loss of gas by leakage or diffusion, whilst still permitting rapid movement of fluid from the clay in about 2 days.

The membrane was previously soaked for one day in the liquid with which the clay was to be subsequently equilibriated. The liquid egress tube and seating for the membrane were filled with liquid and hence any loss of gas through the membrane could be immediately observed. In the infrequent cases when leakage occurred that particular experiment was rejected (Collis-George, 1952). In the case of dehydration experiments, a 5-mm. layer of the clay paste was placed on the wet membrane and the sample allowed to come to equilibrium with a vapour pressure equivalent to the imposed pressure and osmotic pressure before being examined. Five days were found to be sufficient for most specimens to reach equilibrium (constant moisture content), but generally 10 days were allowed.

For hydration experiments the cell was assembled with liquid but without clay and subjected to the required pressure for 1 day to equilibriate the membrane. Then a 1-mm. layer of oven-dry clay was placed on this equilibriated membrane and allowed to hydrate under pressure with excess of liquid (water or 0·45 N NaCl) below the membrane. Again 10 days sufficed for reproducible hydration.

It was found that, provided samples did not touch, a wet clay could be drained and a dry clay wetted simultaneously whilst in equilibrium with the same liquid in one cell, without disturbance of the individual equilibria, which proved to have identical moisture contents with those obtained by separated experiments.

Temperature control, as in the tension plate experiment, was not precise and some experimental error can be expected for this reason.

3. *Vapour pressure measurements* (Thomas, 1928)

The two previous experimental methods avoided the effect of inconstant osmotic pressure upon the total pressure with which the sample is equilibriated as the sample changed moisture content. (This topic is reviewed by Childs and Collis-George, 1950.) In the case of salt-containing clay pastes of arbitrary initial moisture content this requirement is not satisfied using vapour pressure measurements. Vapour pressure measurements were confined, therefore, to wetting and drying salt-free clays and were carried out by the static method in vacuum desiccators evacuated to 2 cm. Hg and with H_2SO_4 solutions at 30·5° C.±0·1. The effect of SO_3 vapour changing the properties of the clay was not considered, the sample generally taking from 5 to 10 days to reach equilibrium. Only at high relative humidities (93·2 per cent.) with energies equivalent to less than 100 atmospheres was the equilibriation unsatisfactory and delayed, perhaps aggravated in part by the chemical reaction of SO_3.

4. *Apparent density measurements*

Specific gravity bottles, originally intended for coal-tar measurements, were used with mercury displacement. The bottles had tapered ground-glass stoppers, but the neck of the bottle was of sensibly the same diameter as its body: the total volume was of the order of 25 c.c. Similar techniques to those used for a Haines's bottle were employed, but before insertion of the stopper a fine capillary was used to free the base and walls of air-bubbles. Measurements were confined principally to samples in equilibrium with an applied pressure and dried from a paste, since cracks and flaws were at a minimum and the total volume of the sample reproducibly determinable. Samples wetted from the ground oven-dry clay had an oolitic appearance and even at low pressures the volume was difficult to determine.

5. *Basal spacing determinations*

A commercial G.E. Corporation X.R.D.3 unit was used with a copper target (40 Kv, 20 ma.) and a nickel filter to produce K_a doublet of wavelength 1·539 Å.

A flat cassette, loaded with 'No-screen' medical film, was used at 5 cm. distance from the specimen. The pressure-equilibriated sample generally was sealed from the atmosphere in a fine pyrex capillary and a powder diagram produced with 45 min. irradiation. For moisture contents less than 25 per cent., attaching particles of the powdered clay to a vaseline-coated thin pyrex rod, or at higher moisture contents using a thin paring of clay with no support, permitted the use of irradiation times of 5–10 min. only. Investigation showed that although basal spacings varied on prolonged exposure of samples to the atmosphere, no variation occurred in the intervals taken for irradiation.

A 4 mm. diameter lead stop was held at the surface of the cassette to eliminate the unscattered primary beam. On the developed films the area around the stop was free of the primary beam and was sufficiently clear to allow measurements of spacings up to 38 Å at 5 cm. from the specimen.

'Basal spacing' in this context is calculated from the (001) reflection. When the clay has a moisture content of greater than 50 per cent., the (00*l*) reflections form a rational series, indicating a regular crystal structure.

Experimental Results

Fig. 1 combines the results for the three pressure techniques. The repression of the diffuse double layer by means of 0·45 N NaCl reduces the water-retaining capacity of the clay relative to salt-free water under all comparable conditions of wetting, and for drying conditions, except at tensions less than 3 cm. water. The hysteresis energy, at any particular moisture content and salt content, represented by the difference of abscissal coordinates in Fig. 1, varies continuously with the moisture content.

In the same way, basal spacing determinations on samples in equilibrium with known pressures show a hysteresis phenomenon, such that at any pressure greater than pF 4 a drying clay has a larger basal spacing than a wetting clay. Below pF 4, for this clay, the salt-free clay is maximally expanded at 19·2 Å whether wetting or drying. The basal spacing of salt-containing clay decreases at pressures lower than for the salt-free clay.

A limited number of absolute values of the basal spacing were found, viz. 19·2, 15·55, 12·55, and 9·85 Å$\pm$0·05, such that on drying a clay with a moisture content equivalent to a basal spacing of 19·2 Å, a succession of combinations of 19·2 and 15·55 Å spacings occur, but no intermediary spacing, the wetter the sample the greater the percentage of the 19·2 Å spacing, and the drier the sample in this range the more of the 15·55 Å spacing, until a stage is reached on drying, when the basal spacing is uniquely 15·55 Å. Further drying produces twin spacings of 15·55 and 12·55; then 12·55 and 9·85; and finally 9·85 Å alone, when oven-dry.

Basal spacings were determined by measurement of the powder diagrams with a vernier scale against an illuminated ground-glass screen, and it was possible, when mixed spacings were encountered, to make some estimate of the fraction of each type of spacing from the density and

thickness of the respective diffraction rings. In Fig. 2, basal spacings for salt-free clay on the right-hand ordinate are plotted against water content. Where twin spacings occurred the 'mean basal spacing' is plotted as a vertical stroke rather than a point: the centre of the line corresponds with this weighted mean and the length of the line to the expected experimental error of the procedure.

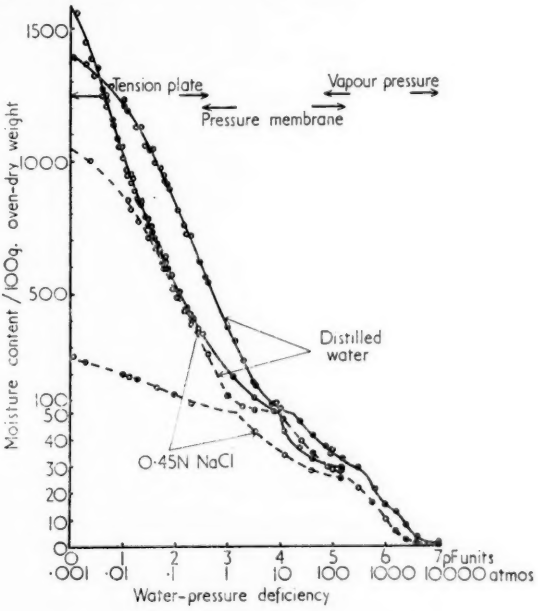


FIG. 1. The moisture characteristics (moisture content–pressure deficiency relationship) of Belle Fourche clay for wetting and drying and in equilibrium with distilled water and 0·45 N NaCl. From 0–50 per cent. moisture content the ordinate has a tenfold magnified scale. The range of pressure covered by each technique is shown. The broken curves refer to the wetting, and the solid curves to the draining procedures described in the text.

The 'mean basal spacing' refers to the spacing which the clay would be expected to have, did the clay but expand in a continuous and non-quantum-like manner, that is, as if the water molecule had no comparatively constant dimensions. Consequently, the distance between similar layers in adjoining crystal sheets could vary continuously and would be identical throughout the whole micelle. It was determined here by weighting the fractions of the constituent twin spacings, themselves calculated from the twin (*ool*) reflections.

These basal spacing results show that after removal of water from 47 per cent. to 32·5 per cent. with progressive decrease from 19·2 to 15·55 Å, further removal of water to 22 per cent. corresponds to a constant

basal spacing of 15·55 Å. Still further dehydration causes a continuous diminution of 'mean basal spacing' to 9·85 Å.

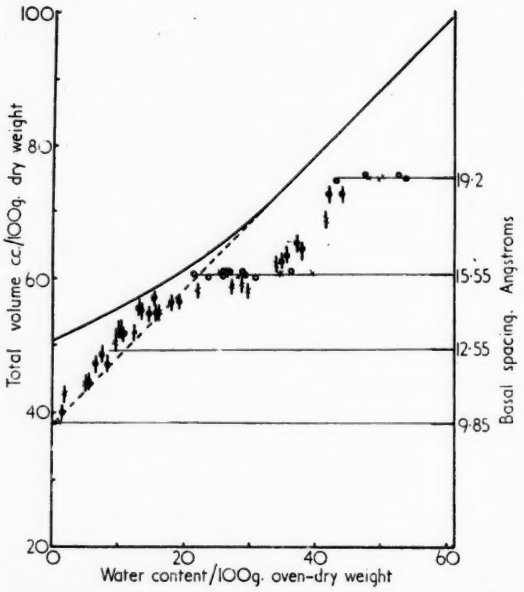


FIG. 2. Specific volume, moisture content, and basal spacing results for Belle Fourche (salt-free). The solid curve denotes the experimentally determined relationship between moisture content and volume on drying. The broken line denotes the theoretical relation between moisture content and volume, assuming no air porosity and density of water equal to unity. The circles refer to dehydration basal spacing results, and the crosses to hydration spacings; the vertical strokes, where present, indicate the presence of two basal spacings, as described in the text.

These results can be further elucidated if the basal spacing data are equated to specific volumes, as on the left-hand ordinate. These are calculated by multiplying the basal spacing by the theoretical cross-sectional area (per 100 g.) of a single montmorillonite crystal, i.e. $3{\cdot}83 \times 10^8$ sq. cm. per 100 g., but see *Discussion* for derivation of this quantity. The specific volume so derived is related to the volume of the solid material of the clay mineral, and to the volume of water (or air) confined between the sheets of the expanding crystal (micelle). That is, this calculated volume is related to internal micellar swelling, as distinct from pores of water (or air) entrapped between micelles which would include water of more than a few monolayers thick on the outside of the micelle.*

* The terms 'intra-micellar' and 'extra-micellar' are here used to describe the space between sheets of the clay crystals which compose the micelle and the space between micelles respectively.

Also in Fig. 2 are shown: (1) theoretical specific volumes for mixtures of clay and water, assuming that the clay has a true S.G. of 2·6, no air is entrapped, and that water has a constant S.G. of unity; and (2) experimentally determined specific volumes for the drying salt-free clay. Although the density measurements were experimentally restricted to drying clay samples, the resultant curve should be reversible under some ideal conditions, e.g. when there is no aggregation, &c. The coincidence of the basal spacing, wetting, and drying results with reference to moisture content is of interest in this respect.

Discussion

For this salt-free clay, Figs. 1 and 2 would suggest that for Na-montmorillonites, and possibly for other monovalent cation clay systems, shrinkage on removal of water proceeds by several recognizable stages, each separated by transitional phases, viz.:

At stage A, zero pressure (1,400 per cent.),* the clay micelles are fully expanded, with a basal spacing of 19·2 Å, but the volume occupied by the micelles (75 c.c.) is small compared with the extra-micellar pores (1,363 c.c.). In fact, although the paste is plastic and not fluid, there is obviously no continuous mineral matrix to confer rigidity, the micelles being presumably randomly distributed and at some distance from each other.

Whilst drying, phase A–B, the clay shrinks by loss of extra-micellar porosity in a strictly linear manner, more work being required, however, to cause each succeeding decrease of volume.

At stage B the micelles just touch to form a rigid framework (47 per cent.) enclosing a much smaller extra-micellar porosity (10·5 per cent.) than heretofore, but retain a maximum basal spacing; the intra-micellar porosity (36·5 per cent.) being large compared with the total specific volume (85·5 c.c./100 g.).

On drying, phase B–C, the clay shrinks by loss of intra-micellar water, the basal spacing being reduced, quantum-like, from 19·2 to 15·55 Å; the extra-micellar porosity remaining constant.

At stage C (32·5 per cent.) the micelles form a rigid skeleton similar to stage B, enclosing the same extra-micellar pore size distribution, but the micelles are now all thinner since there are only two monolayers of water between sheets in the micelle.

During phase C–D the clay shrinks non-linearly, some air enters the extra-micellar pore space, but this extra-micellar porosity nevertheless shrinks as a whole as water is removed, and the spatial configuration of the micelles changes to give a more closely packed structure. The basal spacing remains constant and equivalent to two monolayers of water, whilst the external pores are emptied.

At stage D (22 per cent.) the theoretical specific volume curve and the basal spacing specific volume relationship coincide, the former curve being calculated on the basis of no air porosity and the latter on the basis of no extra-micellar pores. All the water in the clay is in the form of two

* Information in parentheses refers to the results obtained for 100 g. oven-dry Na Belle Fourche Montmorillonite.

monolayers between sheets, the pores being empty and of minimum volume (approx. 4 c.c./100 g.).

During phase D–E the micelles now shrink, quantum-like, from two layers of basal spacing water to one layer, and then from one layer to none. The extra-micellar arrangement, which had coincided with a minimum of porosity at stage D, expands whilst the micelles themselves are contracting, particularly during the removal of the last layer of water from between sheets. During this phase the micelles are larger than would theoretically be derived from calculations using water of density unity between sheets. It must be surmised that either the water molecular arrangement between sheets is more open than in bulk water, or that the water density is unity (or greater) and hence part of the intra-micellar volume is occupied by air. A discussion of the possible molecular arrangements of water (and hence of density) between sheets, which has been given by Barschad (1949), supports a density of water of less than unity.

At stage E (0 per cent.) the basal spacing is a minimum of 9·85 Å, the sheets of the crystal micelle touch along their surfaces and the micelles are arranged to give an extra-micellar (air) pore space as great, or slightly greater, than that at stage B, when the micelles are fully expanded and the pores full of water.

On wetting, under normal conditions, resultant volumes for any moisture content are greater than those drying from a paste, by the amount of air entrapped and not released on addition of water. Nevertheless, apart from a general, but inconstant, increase in extra-micellar porosity by the amount of this trapped air, the basal spacing results indicate that the method of intra-micellar expansions and subsequent extra-micellar swellings will be the reverse of that for shrinkage.

The validity of the basal spacing results to give this shrinkage sequence depends on the validity of two assumptions concerning the density of water and the specific area of clay crystals. The latter quantity is calculable from the dimensions of the clay lattice unit cell, which themselves are calculable from its composition as suggested by MacEwan (1951). The composition given by A.P.I. (1951)* provides values of a_0 and b_0 as 5·16 and 8·93 Å respectively by the method of MacEwan, and hence a specific surface area at maximum dispersion of $76{\cdot}5 \times 10^8$ sq. cm. per 100 g. The total specific volume will be the product of half this specific area (excluding edges) of the clay crystal and the basal spacing. When there is no water in the micelle, this calculation becomes exact, if the calculation of the value of specific surface is also exact; but the greater the number of water monolayers on the surface, and the smaller the number of sheets in a micelle, the larger will be the error of underestimation of the total volume; being of the order of 2·5 per cent. for crystal micelles of twenty sheets having three layers of water between sheets, and on each external surface.

For water contents of less than 50 per cent., which includes stages B to E of the shrinkage mechanism postulated, the water on the outer micelle surfaces will be in equilibrium with pressures of greater than

* $(Al_{1\cdot63}Fe_{0\cdot17}Mg_{0\cdot20})Al_{0\cdot07}Si_{3\cdot93}O_{10}(OH)_2 \cdot Na_{0\cdot24}Ca/2_{0\cdot04}$.

1 atmosphere. Although the theory of Schofield (1946) was devised for thick liquid films, extrapolation of his calculations for mica to these pressures suggests that the layer of water on the outer surfaces is of the order of one monolayer. Hence the estimate of error for three layers of water on all outer surfaces of montmorillonite (a crystal surface similar to mica) should be too large in the range of analysis of Fig. 2.

The derivation of the theoretical specific volume-moisture content relationship with no entrapped air (Fig. 2) assumes that the density of water remains unity. There is little reason to believe that the structural arrangement of water molecules will not vary at clay surfaces, as the basal spacing results have suggested, but until further evidence is produced (see Grim, 1953, for survey to date) the magnitude of this disarrangement can be considered as small because of the observed coincidence of the theoretical and actual volume relationships at moisture contents above 30 per cent.

It is obvious that existing uncertainties as to the exact calculation of specific volumes from basal spacing data, and to the density of water, lead to some ambiguity in the interpretation of results. The apparently satisfactory interpretation, so far given, may have to be more precisely analysed with advances in the knowledge of the variation of water density at surfaces, and of the number of sheets to be found in a micelle.

Experimental confirmation of the mechanism of dehydration postulated, in phase A–B, that is of pore contraction rather than of pore emptying with air entry, was obtained using a benzene-water interface. Benzene-water has a surface tension of 34·42 as compared with water of 71·18 dynes per cm. at 30° C., and hence if the mechanism of dehydration is by pore entry rather than contraction, exertion of a pressure of 1 atmosphere with a benzene-water interface would produce the same moisture content as a pressure of 2·07 atmospheres with air-water. It was found for isolated pressures from zero to 120 atmospheres, that no difference in moisture content was detectable between the two interfaces. Higher pressures were not possible with the 'O' rings and the postulated region of air entry (30 per cent.) at 255 atmospheres air-water could not be reached in the pressure membrane apparatus. However, on applying this technique to Goose Lake illite, a difference of moisture content was found in the postulated air-entry zone; viz. 18·5 per cent. at 120 atmospheres air-water as compared with 14·0 per cent. at 120 atmospheres benzene-water, which is equivalent to 235 atmospheres air-water.

The datum of moisture content has been made arbitrarily that obtained by oven drying at 105° C. The amount of water trapped in the pockets formed by the hexagonal oxygen rings in the surface of the clay has been disregarded in these experiments. The fact that a minimal basal spacing in the presence of 1 per cent. water was obtained is indicative of this arbitrariness. Also, the initial material for 'wetting' experiments was clay dried at 105° C., and the possibility of changes in and between sheets due to this procedure cannot be ignored in a discussion of the hysteresis results of Fig. 1, as by the potential energy model of Everett and Whitton (1952). Mooney, Keenan, and Wood (1952) using 140° C.

as datum, found adsorption was dependent on the initial drying temperature and decided that reproducible adsorption relations could not be obtained for this reason.

The difference between volumes of a clay-water system under a given pressure in equilibrium with different salt concentrations is most marked. By the procedure described, the osmotic pressure contribution to water retention is eliminated and clay with a repressed double layer holds less water than the same clay with a more expanded double layer. More complete data for various ionic concentrations and temperatures is, however, necessary before examination of existing theories on the energy relationships of double layers (e.g. Verwey and Overbeek, 1948) can be attempted.

The determination of moisture content in salty clays is not as precise as that for salt-free clays. For approximate calculation purposes, it can be assumed that all water removed on oven drying is of constant ionic concentration and equal to that of the equilibriating fluid; hence the resultant dry weight can be corrected to a weight of salt-free clay. The assumption is untrue and the error real, but for the results of Fig. 1 it causes an underestimation of less than 1 per cent., as checked by volumetric analysis. The relative position of the curves of Fig. 1 remains unchanged. However, this difficulty prevents the production of specific volume-moisture relationships, for a clay in equilibrium with a salt solution, similar to that of Fig. 2.

In the literature there has been much discussion concerning the hydration of montmorillonite, and this has been recently summarized by Grim (1953) and MacEwan (1951). Consequently, this discussion can be confined to topics directly concerned with the results.

The clay here used, when in equilibrium with water, was not completely sodium-saturated but contained some 13 m.e. per cent. Ca. However, no differences due to this calcium content could be detected between this clay and the same equilibriated with excess NaCl, and certainly none of the spacings peculiar to Ca-clays, as reported in the literature, were found in either system. In the following discussion, therefore, this clay is considered predominantly a Na-clay, but its content of Ca must not be ignored in a final synthesis, when more information becomes available.

Using Na-montmorillonite, Mering (1946) found that diffraction patterns vanished at high water contents. This phenomenon was not observed here, rather the patterns became weakened with increasing dilution, but even at 1,600 per cent. there was a distinct 19·2 Å spacing. However, this provides evidence that the number of sheets per micelle is finite, even when the clay is apparently dispersed, in agreement with Mering's picture of the ultimate crystal unit but not with that of other workers, e.g. Hendricks, Nelson, and Alexander (1940), who consider the ultimate particle to be single sheets 10 Å thick.

For H-montmorillonite, Bradley, Grim, and Clark (1937) found four distinct monolayers of water with increasing hydration. The first two coincide with the 12·55 and 15·55 Å spacings as found here for Na-montmorillonite. The third and fourth monolayers had spacings of 18·4 and 21·4 Å, whereas this clay had only the 19·2 Å spacing. No

difference in spacing, at a given moisture content, was detectable between the clay equilibriated with distilled water or with 0·45 N NaCl. Bradley *et al.* sometimes found intermediary spacings corresponding to statistical averaging of the inter-sheet distances generally present at any one time. This averaging of the primary basal spacing was not found by Mooney *et al.* nor by this author. Averaging of the second and third order basal spacing patterns was sometimes found, and this weighted average was used in analysis of the fractions of the spacings present to confirm the analysis of the first order patterns by optical density examinations.

Mooney *et al.* have shown, by careful vapour pressure experiments, that inflexions in the water content-energy relation correspond to moisture contents of the first and second monolayers of water. Although such critical conclusions are not possible from these results (Fig. 1), it is possible by integration of the area beneath each curve to derive the total energy required or released during a change of moisture content. Converting the integrated values to calories, the heat of wetting and heat of dehydration for the complete range of moisture contents from oven-dry to zero suction calculate respectively to 7 and 14 cals. per g., and the total energy of hysteresis involved in a complete wetting and drying cycle is hence of the order of 7 cals. per g. The energy-moisture relationships are not in fact complete below 1 per cent. in Fig. 1, and these energy values will be underestimates. Provided the moisture content is greater than 1 per cent., theoretical calculations of these work (heat) quantities, required for a change of moisture content, are more convenient and accurate from the energy-moisture curves of Fig. 1 than direct calorimetric determinations. From the shape of the curves it is to be expected that the heat functions will be lower for clays in equilibrium with salt than in salt-free soils, but the lack of a complete energy relation prevents their calculation for a complete cycle.

In conclusion, it appears that the discrepancies between the results of various workers using apparently the same material, e.g., Na-montmorillonite, can only be accounted for by assuming that either the materials were initially different although superficially similar, or that the method of pre-treatment, however slight, has sufficed to permanently modify their intra-micellar structure. Therefore, until more information is available to explain these differences, it behoves every worker to describe precisely the operations and treatments he carried out.

Summary

Complete moisture-energy relationships for a salt-free Na-montmorillonite at 30° C. and partial relationships for the same clay in equilibrium with 0·45 N NaCl were derived by the use of three types of pressure procedures covering different ranges of energy. Specific volumes and basal spacings were determined on clay previously equilibriated with a known pressure.

These results are discussed and a mechanism of clay shrinkage and expansion postulated: this is substantiated in part by the comparison of benzene-water and air-water interfaces. The view of earlier workers that the clay expands or contracts by addition of discrete monolayers of

water is supported, the evidence suggesting that different parts of a micelle can have at the same time different numbers of water monolayers between crystal sheets, i.e. they have, simultaneously, two distinct basal spacings. Contrary to other workers, it was found that Na-montmorillonite had a maximal spacing of 19·2 Å even in very dilute suspensions, i.e. the ultimate crystal is composed of multiple sheets.

The energy relationships allow calculation of the heats of wetting, of dehydration, and of hysteresis.

A plea is entered for the exact description and treatments of the experimental clay materials by authors.

REFERENCES

American Petroleum Institute, 1951. Reference Clay Minerals. Research Project **49**, Columbia University Press, New York.

Barschad, I. 1949. The nature of lattice expansion and its relation to hydration in montmorillonite and vermiculite. Amer. Mineralogist, **34**, 675.

Bradley, W. F., Clark, G. F., and Grim, R. E. 1937. A study of the behaviour of montmorillonite on wetting. Z. Krist. **97**, 216.

Childs, E. C., and Collis-George, N. 1950. The control of soil water. Adv. Agron. **2**, 233.

Collis-George, N. 1952. A note on the pressure plate-membrane apparatus. Soil Sci. **74**, 315.

Everett, D. H., and Whitton, W. I. 1952. A general approach to hysteresis. Trans. Farad. Soc. **48**, 749.

Grim, R. E. 1953. Clay mineralogy. McGraw-Hill, New York.

Haines, W. B. 1930. Studies in physical properties of soils, V. J. Agr. Sci. **20**, 97.

Hendricks, S. B., Nelson, R. A., and Alexander, L. T. 1940. Hydration mechanism of the clay mineral montmorillonite saturated with various ions. J. Amer. Chem. Soc. **62**, 1457.

MacEwan, D. M. C. 1951. The montmorillonite minerals. X-ray identification and structure of clay minerals. Chap. iv. Mineralogical Society of Great Britain Monograph.

Mering, J. 1946. On the hydration of montmorillonite. Trans. Farad. Soc. **42B**, 205.

Mooney, R. W., Keenan, A. G., and Wood, L. A. 1952. Adsorption of water vapour by montmorillonite, I. J. Amer. Chem. Soc. **74**, 1367; II. ibid. 1371.

Richards, L. A. 1947. Pressure membrane apparatus—construction and use. Agric. Engng. **28**, 451.

Schofield, R. K. 1946. Ionic forces in thick films of liquid between charged surfaces. Trans. Farad. Soc. **42B**, 219.

Thomas, M. D. 1928. Aqueous vapour pressure of soils, III. Soil Sci. **25**, 409.

Verwey, E. J. W., and Overbeek, J.Th.G. 1948. Theory of the stability of lyophobic colloids. Elsevier, New York.

THE EFFECT OF BUSH BURNING ON THE MICRO-FLORA OF A KENYA UPLAND SOIL

JANE MEIKLEJOHN

(*East African Agriculture and Forestry Research Organization, Muguga, Kenya*)

THOUGH burning off the vegetation before cultivation is a very common farming practice in the tropics, little is yet known about the effect of this burning on the soil. In the early months of 1953 much new land was burned over on the new experimental farm of the East African Agriculture and Forestry Research Organization; so the opportunity was taken to study some of the changes in the soil microflora produced by burning.

The farm is about 6,800 ft. above sea-level, in the hilly country going up to the eastern escarpment of the Rift Valley; the average annual rainfall is about 36 in. and the soil is a typical red Kikuyu soil, rather heavy, with a high iron content.

The burning of the vegetation on the newly cleared land was carried out during the dry weather between the 'short rains' of November–December, and the 'long rains' of March–May. Three sites, which were burned over on different dates, were studied.

Site B, burned 23 January 1953, was on the east side of a hill, and had been covered with black wattle (*Acacia decurrens*), which is not a native tree, and with a scanty undergrowth of herbs, grass, and small shrubs.

Site C, burned 10 February 1953, had formerly been under African cultivation, but from 1950 it had been allowed to go back to grass; just before burning it was covered with rough grass and some volunteer wattle seedlings.

Site D, burned 12 March 1953, resembled Site B; it was on the west side of the same hill, and was also formerly covered with wattle.

Samples of the top inch of soil were taken from all three sites before burning, and at intervals after burning, and in May 1953 parallel samples were taken from a burnt patch, and a neighbouring unburnt patch, on each site. The numbers of micro-organisms in the samples were counted, using a direct method for total bacteria and actinomycete counts, and plating methods for fungi, bacteria, and actinomycetes separately. Enrichment cultures were made from the samples to determine the presence and degree of activity of nitrogen-fixing and nitrifying bacteria. Table 1 gives the sampling dates, with the rainfall in the intervals between samples.

Methods and Media.

Direct counts of the total numbers of bacteria and actinomycetes were made by the method of Jones and Mollison (1948), in which a small weighed quantity of soil is mixed with melted agar, films of known thickness made from the mixture and stained, and the cells in a number of random fields counted under high power.

Fungi were counted by plating, from a dilution of 1 in 5,000 in 0·75 per cent. saline, on Waksman's acid agar (Waksman, 1922).

Plate counts of bacteria and actinomycetes were made on Thornton's

agar (Thornton, 1922), from a dilution of 1 in a million in saline. The actinomycete colonies were distinguished from bacterial colonies by examination under a lens.

TABLE 1

Sampling Dates and Rainfall

Date	Interval (days)	Rainfall in interval (in.)	Samples taken from sites		
			B	*C*	*D*
1 Jan.	..	..	..	..	..
13 Jan.	13	1·32	..	..	..
21 Jan.	8	nil	B2	..	..
23 Jan.	2	nil	Burnt, B3	..	..
27 Jan.	4	nil	B4	..	..
4 Feb.	8	nil	B5	..	..
10 Feb.	6	nil	..	C1, Burnt	..
12 Feb.	2	nil	..	C2	..
19 Feb.	7	nil	B6	..	..
20 Feb.	1	nil	..	C3	..
24 Feb.	4	nil	B7	..	..
4 Mar.	8	0·37	..	C4	..
12 Mar.	8	0·14	..	..	D1, Burnt
13 Mar.	1	nil	..	..	D2
17 Mar.	4	nil	..	C5	..
20 Mar.	3	0·03	..	..	D3
9 April	20	1·12	All sampled for nitrogen		
15 April	6	3·03	..	..	D4
13 May	30	6·97	B8, B9	..	..
15 May	2	0·08	..	..	D5, D6
21 May	5	0·06	..	C6, C7	..

Aerobic nitrogen-fixers; a 1 in 10 dilution of soil in saline was streaked on a plate of Jensen's no-nitrogen agar (Jensen, 1951*a*). The growth developing on the plate was picked off on to slopes of dextrose-chalk agar (agar 15 g., dextrose 20 g., calcium carbonate 20 g., magnesium sulphate 0·5 g., tap water 1,000 ml.).

Anaerobic nitrogen-fixers were obtained by enrichment cultures on potato (Jensen, 1951*b*); positive cultures, which contained rods staining blue with iodine, were transferred to narrow tubes of semi-solid Jensen's agar (0·1 per cent. agar).

Nitrifiers were obtained in enrichment culture by incubating 2 g. soil in 40 ml. liquid medium in a 250-ml. conical flask. The media contained either 0·005 M ammonium sulphate or 0·0072 M sodium nitrite, and also 0·005 M sodium chloride, 0·001 M magnesium sulphate, 0·001 M potassium dihydrogen phosphate, 0·0001 M ferrous sulphate, 0·00002 M copper sulphate, and 1·0 g. calcium carbonate per litre, made up in tap water (Meiklejohn, 1953).

Results

The results of the counts on the three sites are given in Tables 2, 3, and 4. The *direct counts* show a drop immediately after burning in all three cases. This method of counting does not distinguish between

TABLE 2

Site B: Numbers of Organisms per gramme of Dry Soil

Sample	*Date*	*Days after burning*	*Direct count (millions)*	*Plate counts*		
				Fungi (thousands)	*Actinomycetes (millions)*	*Bacteria*
B2	21/1	(2 before)	2,219	23	157	340
	Burnt 23/1/53					
B3	23/1	0	1,807	4·3	10·3	35·4
B4	27/1	4	2,658	0	26	19·7
B5	4/2	12	2,737	0	12·5	27
B6	19/2	27	2,310	15·8	..	..
B7	24/2	32	2,115	..	30	10
B8	13/5 (burnt soil)	4 months	1,642	21·4	155	49
B9	13/5 (unburnt soil)	4 months	2,607	30	130	44·5

(Plate counts the average of 4 plates)

TABLE 3

Site C: Numbers of Organisms per gramme of Dry Soil

Sample	*Date*	*Days after burning*	*Direct count (millions)*	*Plate counts*		
				Fungi (thousands)	*Actinomycetes (millions)*	*Bacteria*
C1	10/2	(before)	3,123	25	520	43
	Burnt 10/2/53					
C2	12/2	2	1,945	5·4	236	33·5
C3	20/2	10	2,210	0	257	25
C4	4/3	22	2,297	2·9	73	16
C5	17/3	35	1,074	29·5	236	4·8
C6	21/5 (burnt soil)	3 months	1,200	34·8	39	38
C7	21/5 (unburnt soil)	3 months	1,856	18·9	31·6	12·2

(Plate counts the average of 4 plates)

TABLE 4

Site D: Numbers of Organisms per gramme of Dry Soil

Sample	*Date*	*Days after burning*	*Direct count (millions)*	*Plate counts*		
				Fungi (thousands)	*Actinomycetes (millions)*	*Bacteria*
D1	12/3	(before)	2,820	37·4	392	13·8
	Burnt 12/3/53					
D2	13/3	1	1,974	14·6	183	5·4
D3	20/3	8	1,983	5·2	292	6·3
D4	15/4	34	2,865	24·6	251	23·8
D5	15/5 (burnt soil)	2 months	1,355	19	131	22
D6	15/5 (unburnt soil)	2 months	2,460	13	36	10·6

(Plate counts the average of 4 plates)

actinomycete spores and bacterial spores, or spherical bacteria (cocci). It is usually considered to overestimate the total number of organisms present, as dead cells may be included in the count. The numbers estimated by this method are very high, of the order of thousands of millions per gramme of soil, and it is interesting to note that numbers of the same order of magnitude were obtained in counts made by the same method on Broadbalk Field at Rothamsted (Skinner, Jones, and Mollison, 1952). After the initial fall in numbers due to burning, there was a rapid recovery on Site B, where the direct count was back to its original level 4 days after the burn. On Sites C and D recovery was very slow; the reason for this was probably the drought from mid-January to mid-March (Table 1). The count on Site D on 15 April (D4), which shows recovery to the original level, was made the day after 1·75 in. of rain during the night. The direct counts made in May show higher numbers in the unburnt patches than the burnt ones.

Fungi, as estimated by the plate count, were completely eliminated by burning on Sites B and C, and much reduced in numbers on Site D. Burning did not, however, appear to alter the composition of the fungus population on Sites B and C; in both cases the dominant organism before burning was a species of *Penicillium* with compact colonies, green above and white or yellow below; and the same species was still dominant in the fungus flora when it was re-established after burning (Table 5). On Site D, however, the same *Penicillium*, which was the dominant form before burning, was replaced by an olive-green *Aspergillus* after the burn; the same *Aspergillus* was also dominant in the sample taken in May from unburnt soil on Site B. *Trichoderma* spp. were present on Sites B and D, both of which had carried trees, but not on Site C, which had been under grass (Table 5).

Plate counts of bacteria and actinomycetes underestimate the total population very considerably; Thornton's medium, which was used for these counts, allows the development of a great variety of organisms, but there are species which will not grow on it. Counts by the plating method accordingly give figures tenfold or even a hundredfold less than the direct count (Skinner, Jones, and Mollison, 1952). But, like the direct counts, the plate counts of both bacteria and actinomycetes show a fall after burning on all three sites. At the May observation, however, the plate counts were higher on the burnt soil than on the unburnt from all three sites. On Sites C and D the numbers of actinomycetes were consistently greater than the numbers of bacteria.

Nitrogen-fixers. *Azotobacter*, the classical aerobic nitrogen-fixing bacterium, was never found in any of the samples. The only cultures obtained which were found to fix appreciable quantities of nitrogen aerobically were found in Sites B and D before burning. A sample from B taken on 16 December 1952 yielded a culture of a small rod-shaped bacterium which fixed 1·22 mg. nitrogen in 10 ml. culture in 14 days. Two cultures were obtained from Site D on 12 March, just before burning, which fixed 1·85 mg. and 4·85 mg. nitrogen in 10 ml. in 14 days. The samples taken from B after burning all gave negative results; twenty aerobic cultures were examined, but none could fix nitrogen. It there-

TABLE 5

Genera of Fungi Identified on Plates

Site B		*Site C*		*Site D*	
Sample	*Number of colonies*	*Sample*	*Number of colonies*	*Sample*	*Number of colonies*
B1	1 Penicillium 1 Fusarium 3 unidentified	C1	9 Penicillium 3 Verticillium 1 Aspergillus 5 unidentified	D1	14 Penicillium 4 Fusarium 4 Monosporium 2 Aspergillus 2 Trichoderma 1 Monilia 2 unidentified
B2	4 Penicillium 2 Verticillium 11 unidentified		Site C Burnt		Site D Burnt
	Site B Burnt	C2	1 Penicillium 1 Mucor 2 unidentified	D2	4 Aspergillus 3 Penicillium 2 Mucor 1 Monosporium 1 unidentified
B3	1 Penicillium 1 Aspergillus 1 unidentified	C3	Nil		
		C4	2 unidentified		
B4	Nil	C5	5 Penicillium 2 Verticillium 1 Monosporium 1 Aspergillus 1 Mucor 3 unidentified	D3	4 Aspergillus 1 Mucor
B5	Nil			D4	7 Aspergillus 5 Penicillium 1 Fusarium 2 unidentified
B6	1 Penicillium 11 unidentified				
B8	(Burnt soil) 6 Penicillium 3 Trichoderma 2 Pythium 1 Aspergillus 1 Monilia 2 unidentified	C6	(Burnt soil) 22 Penicillium 3 Mucor 2 Monosporium 1 Fusarium 1 Aspergillus	D5	(Burnt soil) 5 Penicillium 3 Mucor 2 Aspergillus 1 Monilia 2 unidentified
B9	(Unburnt soil) 13 Aspergillus 2 Trichoderma 2 Verticillium 3 unidentified	C7	(Unburnt soil) 9 Penicillium 2 Mucor 1 Aspergillus 1 Spicaria 2 unidentified	D6	(Unburnt soil) 4 Mucor 2 Trichoderma 1 Aspergillus 1 Penicillium 1 Fusarium

fore appears that the aerobic nitrogen-fixers were killed by burning. Site C yielded a rather poor culture before burning (0·87 mg. fixed in 10 ml. in 14 days), and two doubtful nitrogen-fixers after burning, one from the sample taken 2 days, the other from the sample taken 10 days later. Site D yielded a doubtful culture 8 days after burning, but no other culture from either C or D could fix nitrogen under aerobic conditions.

Clostridium species, which fixed small quantities of nitrogen under anaerobic conditions, were obtained from Site B immediately, 4 days, and 32 days after burning; from Site C 10 and 22 days after burning, and from D 34 days after burning. The quantities of nitrogen fixed were very small, the largest being 0·48 mg. in 10 ml. in 14 days, from a Site-B culture. It therefore appears that the anaerobic nitrogen-fixers, though not very effective, do survive bush burning.

Nitrifiers. Enrichment cultures which contained both ammonia-oxidizing and nitrite-oxidizing bacteria were obtained from all three

sites before burning. They were not obtained in pure culture, but the appearance of cells in the cultures was similar to cultures of the classical autotrophic nitrifiers, *Nitrosomonas* and *Nitrobacter*. On Site B the nitrifiers were apparently killed by burning, as no oxidation took place in media inoculated with soil collected from B 4, 12, and 27 days after burning. The cultures made from Sites C and D after burning were all very slow-acting, so that it is probable that only a few nitrifiers survived the burn.

Nitrogen determinations were made in the Chemistry Department on parallel soil samples taken on 9 April, from the surface of burnt and unburnt patches on all three sites. The results, which are given in Table 6, show that the total nitrogen content was higher in the burnt soil than in the unburnt, on all three sites. This was 11 weeks after burning on Site B, 8 weeks on Site C, and 4 on Site D. The sampling was repeated on 18 August, 5 months later, and the results of burning were still perceptible, as once again there was more nitrogen in the burnt soil on all three sites.

TABLE 6

Total Nitrogen in Burnt and Unburnt Soil

	9 April	*18 August*
Sample	*Nitrogen % of dry soil*	*Nitrogen % of dry soil*
B: Burnt . .	0·73	0·935
B: Unburnt .	0·40	0·345
C: Burnt . .	0·66	0·490
C: Unburnt .	0·43	0·445
D: Burnt . .	0·77	0·600
D: Unburnt .	0·57	0·525

Discussion

The practice of bush burning is regarded with a cautious eye by some Agricultural Departments in tropical countries, as it is thought that the clearing of ground cover and the destruction of roots in the top soil may lead to erosion. There are, however, few recorded measurements of the extent of erosion which may follow burning. Trapnell[1] found that 11 years of repeated burning caused slight erosion on a site in Northern Rhodesia with a slope of only 3 per cent. The effect of burning on the soil is probably confined to the top soil, as Masson (1948), in a very careful study of bush burning in Senegal, found that the greatest rise of temperature recorded by a thermocouple buried 2 cm. under the surface of the soil under a bush fire was only 14·4° C. On the surface of the soil, however, a fire burning fiercely in bush 4 ft. high, with a light wind, raised the temperature momentarily to 850° C.

As against the risk of erosion, however, must be set the fact that better yields are obtained after burning in many places. Controlled burning, in the dry season before the rains, gives better pastures in parts

[1] Private communication.

of East Africa, because it destroys herbs and clears away the mat of old dry tussocky grass, and so encourages the development of more succulent and palatable types of grass (Cook, 1939; Edwards, 1942; van Rensburg, 1952).

It is also known that increased yields of arable crops often follow burning, but the reasons for this are not yet fully understood. On the Muguga farm a better yield of beans was obtained in 1952 from land which had been burnt over than from adjacent unburnt land. Mr. Ledger, the Farm Manager, considers that this was partly due to the destruction of weeds, and of crop residues, by burning. Young and Golledge (1948), working in Northern Rhodesia, found that the wood-ash formed by burning off trees and shrubs supplied the soil with calcium, phosphorus, potassium, magnesium, and sulphur, in all of which the local soil was seriously deficient; these elements were drawn up by the trees from the subsoil, concentrated in the tissues of the tree, and released again into the topsoil in the wood-ash. Corbet (1934) suggested that wood-ash, as well as supplying potash to the soil, partly neutralized acid soils.

Another possible reason for the better yields after bush burning is suggested by the present work; there was more total nitrogen in the burnt soil than in adjacent unburnt patches on all three sites. Unfortunately it was not possible to determine in what forms this nitrogen was present, so that it cannot be known how much of it would have been available to a crop. The reason for the greater total nitrogen in the burnt soils may have been the temporary destruction of the soil fungi; but it is more likely to have been the suppression of the nitrifying bacteria after burning.

Nitrate, which is formed by the nitrifying bacteria, is easily lost by denitrification or by leaching, so that suppression of nitrification could lead to conservation of nitrogen in the top soil. The rainfall in March and April 1953 was, however, very small (Table 1), so that losses by leaching are unlikely.

It is interesting that a reduction in the number of micro-organisms, whether estimated by direct counting or the plating method, was found to follow burning on all three sites. Corbet (1934), in a study of cleared forest in Malaya, found a rise after burning in the total number of colonies (fungi, bacteria, and actinomycetes) developing on plates of Thornton's agar, in one case from 500,000 to 5 million cells per gramme moist soil, and in another case from 500,000 to 11 million. He attributed the rise in numbers to the effects of the addition of wood-ash to the soil. It is evident, however, that plate counts on any medium only represent a fraction of the total soil microflora; the numbers estimated by direct counting are so very much greater, ten or even a hundred times the estimates of numbers obtained by plating.

Summary

1. Soil samples were taken from the surface of three sites at Muguga, Kenya, before and at intervals after the vegetation covering the sites was burnt.

2. The numbers of micro-organisms estimated by a direct-counting method fell after burning on all three sites.
3. There was also a reduction after burning in the numbers estimated by plate counts, especially in the number of fungi, which disappeared completely for a short time on two of the sites.
4. Aerobic nitrogen-fixers are apparently killed by burning; but anaerobic nitrogen-fixers, though not very efficient at any time, appear to survive burning.
5. The nitrifying bacteria are either killed, or reduced to very few, by burning.

REFERENCES

COOK, L. 1939. A contribution to our information on grass burning. S. Afric. J. Sci. **36**, 270.

CORBET, A. S. 1934. The bacterial numbers in the soil of the Malay peninsula. Soil Sci. **38**, 407.

EDWARDS, D. C. 1942. Grass-burning. Emp. J. Expt. Agric. **10**, 219.

JENSEN, H. L. 1951*a*. Notes on the biology of Azotobacter. Proc. Soc. Appl. Bact. **14**, 89.

—— 1951*b*. Notes on the microbiology of soil from Northern Greenland. Medd. Grønland, **142**, 23.

JONES, P. C. T., and MOLLISON, J. E. 1948. A technique for the quantitative estimation of soil micro-organisms. J. Gen. Microbiol. **2**, 54.

MASSON, H. 1948. La température du sol au cours d'un feu de brousse au Sénégal. Agron. Trop., Nogent, **3**, 174.

MEIKLEJOHN, J. 1953. Iron and the nitrifying bacteria. J. Gen. Microbiol. **8**, 58.

SKINNER, F. A., JONES, P. C. T., and MOLLISON, J. E. 1952. A comparison of a direct- and a plate-counting technique for the quantitative estimation of soil micro-organisms. J. Gen. Microbiol. **6**, 261.

THORNTON, H. G. 1922. The development of a standardised agar medium for counting soil bacteria. Ann. Appl. Biol. **9**, 241.

VAN RENSBURG, H. J. 1952. Grass burning experiments on the Msima River stock farm. East Afric. Agric. J. **17**, 119.

WAKSMAN, S. A. 1922. A method for counting the number of fungi in the soil. J. Bact. **7**, 339.

YOUNG, R. S., and GOLLEDGE, A. 1948. Composition of woodland soils and wood ash in Northern Rhodesia. Emp. J. Expt. Agric. **16**, 76.

MICROBIOLOGICAL STUDIES OF SOME SUBANTARCTIC SOILS

J. S. BUNT* AND A. D. ROVIRA†

(*Microbiology Laboratories, School of Agriculture, University of Sydney.*)

Summary

A study has been made of the rhizosphere effect of four subantarctic plant associations at Macquarie Island. A stimulation occurred in each case in the rhizosphere, even in the more primitive *Azorella selago* and *Dicranoweisia antarctica* associations.

No qualitative differences were found between rhizosphere and non-rhizosphere isolates. The temperature range of most isolates was 10° to 25° C. with a high proportion giving maximum growth at 25° C.

Respiration measurements were made on 11 soils using the Warburg respirometer and a marked correlation was found between total carbon content of the soil and respiration. The variations in respiratory quotient which occurred were not related to carbon content and were probably due to differences in chemical constitution of the organic-matter fractions and soil aeration.

Introduction

THE islands of the subantarctic represent a zone in which soil development and the growth of vascular plants are severely limited or, in some areas, completely prevented by climatic conditions. Considerable interest should be attached to a study of the nature and activity of the microflora of such soils. This investigation suggested itself as an extension of a general microbiological survey conducted by one of us (Bunt) on the soils of Macquarie Island whilst a member of the 1951–2 Australian National Research Expedition. It is concerned with the uptake of oxygen and evolution of carbon dioxide by these soils compared with a red loam from the Australian mainland and also with the influence of certain subantarctic plant alliances on the soil microflora.

Macquarie Island, situated in 54° 30′ S. lat. and 158° 57′ E. long., lies at a point approximately midway between Tasmania and the nearest point of the Antarctic continent. Twenty miles long and less than 4 miles wide, its chief topographic features are the raised beach terrace which is generally very narrow, but up to half a mile wide over the north-west sector of the coastline, and the steep cliff areas rising to the more or less undulating plateau some 800–1,000 ft. above sea-level. The geological structure of the island (Mawson, 1943) is volcanic.

The climate of Macquarie Island is severe, with a mean temperature of 4·4° C. and a mean annual range in temperature of about 2·8° C. The mean annual wind velocity is approximately 15 miles per hour. A mean of only 1·1 hours of sunshine is recorded per day. The mean humidity for each month is 90 per cent. Precipitation in the form of rain, hail, sleet, mist, and snow amounts to 45 in. per annum. There is no permanent cover of snow or ice anywhere on the island. A full account of the

* Biologist, Australian National Antarctic Research Expedition.

† Present address: Institute of Microbiology, Royal Agricultural College, Uppsala, Sweden.

meteorological findings may be found in the *A.N.A.R.E. Reports*, Series D, vols. i–iv, 1948–51.

Because of the harsh conditions and the distances which separate the island from other land masses, the vascular flora is scant (Cheeseman, 1919; Taylor, in press), consisting of 38 species of grasses and herbs. Most species occur around the narrow coast line and on the lower slopes leading up to the plateau where the exposure to wind is least.

The scant vegetation of the higher parts of the plateau is dominated by a button moss, *Dicranoweisia antarctica* (C. M.) Par. and the large wind-resistant cushions of *Azorella selago* Hook., which latter afford some protection for the growth of several other species, notably *Pleurophyllum Hookeri* Buch. and *Stilbocarpa polaris* A. Gray. The 'soil' in these areas consists of scarcely altered morainal debris or till, often many feet thick. Beneath the *Azorella* cushions is a well-defined zone of dark, partly decomposed organic material some 12 in. thick.

The plant cover of the lower parts of the plateau is more or less complete. The principal species are *Pleurophyllum Hookeri*, *Stilbocarpa polaris*, and *Azorella selago*, and these have given rise to a characteristic dark peaty soil several feet thick.

The less precipitous slopes rising from the coast are densely and heavily covered with vegetation, the dominant species being *Poa foliosa* Hook f., a massive tussock-forming grass, and *Stilbocarpa polaris*, the latter being stunted on the plateau, but under these conditions luxuriant and leafy. This association produces a wet red-brown peat up to 5 ft. thick, usually lying directly on the parent rock. These plants also dominate the narrow coastal belt, forming large raised humps of old but little altered root masses. Between these, there are holes commonly several feet deep, filled with a foul-smelling mud or watery slime. Where underlying beach sands provide free drainage, the spaces between the tussocks are dry or only slightly wet. *Pleurophyllum Hookeri–Stilbocarpa polaris* associations displace the tussock on the wide raised terrace of the north-west coast and overlie a layer of completely waterlogged bog peat some 15 ft. thick.

In addition to these formations there are the maritime communities that have colonized the rocky areas jutting into the sea and the beach areas close to the shoreline. These pave the way for succession by less hardy forms.

Soils Studied

Sample numbers 1A, 1B, 1C. Surface 6 in. of light brown gravelly morainal mud free of vegetation within the *Dicranoweisia antarctica* associations.

Sample numbers 1D, 1E, 1F. The material immediately beneath and adhering to the underside of the buttons of the moss *Dicranoweisia antarctica*. A grey-brown gravelly mud containing small quantities of organic matter.

Sample numbers 2A, 2B, 2C. Surface 6 in. Similar to 1A, 1B, 1C, but within the *Azorella selago* associations.

Sample numbers 2G, 2H, 2I. The dark brown, partly decomposed organic material immediately surrounding the roots and densely packed subaerial stems of *Azorella selago*.

Sample numbers 2J, 2K, 2L. Dark brown peaty samples from the zones of the vigorous fleshy main root system of *Pleurophyllum Hookeri* growing in very close association with *Azorella selago.*

Sample 3. A composite sample from the *Azorella selago–Pleurophyllum Hookeri* alliance above.

Sample 24. Sandy material of basaltic origin from the root mass of beach tussock, *Poa foliosa,* growing about 20–30 yds. from the water's edge in a well-drained situation much frequented by elephant seals.

Sample numbers 4A, 4B, 4C. Surface 6 in. of coarse basaltic beach sand from bare areas between clumps of *Cotula plumosa,* Hook. f., immediately adjacent to the inter-tidal zone.

Sample numbers 4D, 4E, 4F. Sand thickly penetrated by the roots of *Cotula plumosa.* This area is also frequented by seals.

Sample number 32. Sandy material bound by the roots of a small grass, *Puccinellia macquariensis* Cheesem., which grows in dense tufts on rock piles jutting into the sea and is constantly exposed to spray. This is one of the plants which prepares the way for succession by the prolific *Poa foliosa.*

Sample number 38. Highly organic, dark brown, partly decomposed peaty material from the surface 6 in. between tussocks in the extensive *Poa foliosa–Stilbocarpa polaris* alliances occupying the steep slopes leading up to the plateau.

Experimental

Sampling

Triplicate samples of rhizosphere and non-rhizosphere soils were collected and placed in sterile bottles in April 1953. Immediately after returning to Australia (a voyage of 7 days) the samples were transferred to a cool room maintained at 5° C. Three weeks elapsed between sampling and the commencement of the experiments.

Where, as in the *Dicranoweisia antarctica* and *Azorella selago* associations, no distinction could be made between true roots and the buried vegetative portions, the rhizosphere was considered as the soil that adhered to any buried living plant material. For the remaining two associations, the rhizosphere was considered as that soil which adhered firmly to the roots.

Plate count estimation of numbers of bacteria

A modified soil extract agar medium of the following composition was used:

$K_2H\,PO_4$	0·4 g.	Peptone	1·0 g.
$(NH_4)_2\,HPO_4$	0·5 g.	Yeast extract*	1·0 g.
$MgSO_4.7H_2O$	0·05 g.	Soil extract†	250 g.
$Mg\,Cl_2$	0·1 g.	Tap water	750 g.
$Fe\,Cl_3$	0·01 g.	Agar	15 g.
$CaCl_2$	0·1 g.		

pH to 7·4, steam for 30 min., and filter.
Autoclave 10 lb. for 20 min.

* Yeast extract consisted of 'Vegemite', a commercial product.

† Soil extract—prepared by autoclaving 1 kg. soil with 1 litre H_2O for 15 min. and filtering.

Duplicate plates of each dilution were incubated at 25° C. for 14 days. The effect of temperature on plate count was studied for four samples by incubating duplicate sets at 10° C. for the same time.

The number of micro-organisms in each rhizosphere sample was estimated by suspending the roots and firmly adhering soil in 100 ml. sterile water, shaking, preparing tenfold dilutions, and plating with the above medium. The dry weight of root material and soil in each sample was determined for subsequent expression of results.

Results are per gramme of oven-dry soil and, additionally, rhizosphere counts are given per gramme of dried root material.

Qualitative tests of bacteria

The following characteristics were studied by the usual methods:

Morphology.
Gram-stain reaction.
Reduction of nitrate to nitrite.
Production of ammonia.
Proteolytic activity.

Temperature requirements of isolates

The optimum temperature of growth of isolates from plates incubated at 10° and 25° C. was assessed by comparing the turbidity of soil extract-yeast extract-peptone broth cultures incubated at the temperatures 10°, 25°, and 37° C.

Organic-matter analyses

Organic-carbon content of each soil was determined by the rapid method of Walkley–Black (1947) and nitrogen determinations were made by the Kjeldahl technique.

Soil respiration

The O_2 uptake and CO_2 evolution were determined by means of the Warburg respirometer (Rovira, 1953). Duplicate sub-samples, each of 7 g. wet weight, were used on each soil for both O_2 uptake and CO_2 evolution, measurements being taken at 25° C. over a period of 2 hours.

Results

Carbon-nitrogen values

The increase in organic matter in the rhizosphere samples (Table 1) is to be expected since the plants grow either in tussocks (*Poa foliosa* and *Cotula plumosa*) or in terraces (*Azorella selago* and *Pleurophyllum Hookeri*). The organic materials formed from past generations of plants are largely held in place by the habit of growth of the living species so that the bare ground between the clumps remains highly mineral in composition.

The low organic status of soils 3A and 24 may be attributed to the fact that they occur in a well-drained area just beyond the littoral, resulting in poor conditions for plant growth and hence for soil development.

TABLE 1

pH, *Total Carbon, Total Nitrogen, and Certain Other Characteristics of Some Macquarie Island Soils*

Sample no.	pH (*Feb. 1952*)	*Total carbon** %	*Total nitrogen** %	C/N	*Elevation above sea-level (ft.)*	*Temperature (° C.) of soil recorded in Feb.*	*Moisture content % (Feb. 1952)**
1A	6·9	2·1	0·22	9·5	800	4·8	25·2
1D (rhizosphere)	6·8	3·6	0·30	12·0	800	3·8	39·3
2A	6·2	3·2	0·21	15·2	700	5·0	38·6
2J (rhizosphere)	6·2	12·5	0·78	16·0	700	4·8	248·0†
3A‡	4·6–4·8	0·7	0·16	4·4	10	10·0	15·1
24‡ (rhizosphere)	4·3–7·0	1·3	0·11	11·8	10	8·0	37·7
32 (rhizosphere)	6·7	4·4	0·69	6·4	10–20	8·2	81·8
38	4·0	43·6	3·33	13·1	100	4·8	675·0†

* Expressed on oven-dry basis.
† Waterlogged peats with very high absorption capacity. Certain semi-fluid bog peats on Macquarie Island contain only 2 per cent. solid matter.
‡ 3A and 24 are closely related to 4A–C and 4D–F respectively, for which figures are not available.

In Table 2 the results of plate counts at 10° and 25° C. and the mass of soil and plant material in each sample are given.

An analysis of the plate counts obtained at 25° C. and expressed as millions per gramme of soil showed that there is a highly significant ($P < 0{\cdot}01$) increase in the population of the rhizosphere, and that the differences in populations between soils are also significant.

The above results show a greater variation than has been obtained by Rovira (unpublished results) with samples of a single species plant-soil association at Lismore, New South Wales. This was probably caused by the gravelly or rocky nature of some of the soils and the rather unusual growth habit of several of the plant species used in this study. Both factors made it difficult to achieve a uniform sampling technique.

The plate counts at 25° C. do not differ significantly from the replicate plates incubated at 10° C. which approaches the mean soil temperature. A study of temperature range and optimum temperatures for growth of isolates picked at random from plates incubated at both temperatures showed no differences due to isolation temperature. These results are reported in Table 3. They indicate that the organisms vary in their optimum temperature requirements but all grow in the temperature range of 10° to 25° C.

A study of the qualitative nature of 40 rhizosphere and 40 non-rhizosphere isolates was made and the results are shown in Table 4. No differences in biochemical activity between the two groups of isolates could be demonstrated.

In Table 5 are presented the figures in μl./g./hour* at 25° C. for oxygen uptake, carbon-dioxide evolution, and associated R.Q. for 11 Macquarie Island soils, and a soil from Lismore (subtropical region) which was included for purposes of comparison. The experiment was conducted at 25° C., which was found to be the optimum temperature for the growth of most organisms isolated from the above soils on soil-extract agar. The results of experiments in which the respiration of these

* Oven-dry weight.

soils was studied over a range of temperatures from 10° to 52° C. will be presented elsewhere.

TABLE 2

Numbers of Micro-organisms in Rhizosphere and Non-Rhizosphere Samples

Sample	Mass soil (g.)	Mass root (g.)	Plate count $\times 10^6$ at 25° C.	
			g. soil	g. root
1. Non-rhizosphere				
(A)	10·45	..	0·44	..
(B)	10·23	..	0·24	..
(C)	12·36	..	0·83	..
1. Rhizosphere				
(D)	8·71	0·63	1·00	12·8
(E)	7·33	0·92	1·00	7·75
(F)	6·51	0·41	1·41	22·6
2. Non-rhizosphere				
(A)	10·15	..	0·46(0·43)*	..
(B)	4·57	..	0·54	..
(C)	6·21	..	1·49	..
2. Rhizosphere				
(G)	1·98	0·71	2·45(1·67)	6·80(4·81)
(H)	3·62	0·84	10·8	46·4
(I)	3·41	0·42	6·49	52·6
2. Rhizosphere				
(J)	6·06	0·09	2·72	183·0
(K)	4·30	0·20	3·35	73·5
4. Non-rhizosphere				
(A)	14·37	..	2·71(2·89)	..
(B)	18·13	..	5·09	..
(C)	17·80	..	2·53	..
4. Rhizosphere				
(D)	17·46	0·09	14·0(14·2)	2,730(2,870)
(E)	16·03	0·07	6·82	1,580
(F)	14·82	0·08	15·6	2,490

* Numbers in brackets represent comparable plate counts at 10° C.

TABLE 3

Temperature Requirements of Isolates from Macquarie Island Soils

Soil	Temperature of isolation	Number of isolates	% isolates giving maximum growth at:			% isolates with equal growth at:	
			10° C.	25° C.	37° C.	10°+ 25° C.	25°+ 37° C.
2(A)	10° C.	20	15	65	0	20	0
(G)*	10° C.	19	21	53	0	26	0
4(A)	10° C.	18	0	100	0	0	0
(D)*	10° C.	20	0	70	0	30	0
2(A)	25° C.	18	55	28	0	17	0
(G)*	25° C.	18	6	61	0	33	0
4(A)	25° C.	17	6	82	0	6	6
(D)*	25° C.	19	5	84	0	11	0

* Rhizosphere samples.

TABLE 4

Morphology, Gram Reaction, and Activity of Isolates

Soil	Number of isolates	Gram stain			Morphology			Gelatin liquefaction		Ammonia production		Nitrate reduction	
		+	−	+/−	Rods	Coccoid rods	Cocci	+	−	+	−	+	−
2(A)	20	4	10	6	19	1	0	13	7	2	18	4	16
(G)*	20	9	7	4	13	2	5	2	18	5	15	5	15
4(A)	20	6	13	1	15	5	0	4	16	7	13	12	8
(D)*	20	2	16	2	18	2	0	6	14	5	15	6	14

* Rhizosphere samples.

TABLE 5

Uptake of O_2 *and evolution of* CO_2 *in* $\mu l./g./hr.$ *at* 25° C.

Sample number	Soil type and plant cover	CO_2	O_2	R.Q.
1D	Glacial till—moss cover	6·5	5·0	1·30
1A	Glacial till—bare	5·5	4·6	1·20
2J	Glacial till—*Azorella selago* cover	20·0	21·0	0·95
2A	Glacial till—bare	3·6	1·9	1·90
19	Peat between *Poa foliosa* tussocks on coast	24·0	37·0	0·65
23	Peat-upland sub-glacial herb field	85·0	65·0	1·31
24	*Poa foliosa* root mass in beach sand	3·0	1·5	2·00
3A	Beach sand close to No. 24	1·7	0·8	2·12
32	*Puccinellia macquariensis* root mass	20·0	45·0	0·44
38	Peat between *Poa foliosa* tussocks on steep slopes	130·0	191·0	0·63
43	Peat-lowland sub-glacial herb field	120·0	96·0	1·24
	Lismore red loam	5·0	3·5	1·43

The most obvious fact shown by these results is the wide variation both in oxygen uptake and carbon-dioxide evolution and in R.Q. amongst the soils examined. That the respiration should vary to such an extent is not surprising when one considers that the soils examined ranged from almost pure beach sand of basaltic origin through several types of more or less primitive organic soils to morainal debris barely showing signs of profile development. It must also be remembered that the samples were taken at a range of elevations from sea-level to about 1,000 ft. within a horizontal distance of less than a mile, and that a number of quite distinct ecological environments have been represented. A striking difference in respiration was exhibited in a *Pleurophyllum Hookeri* environment (soil 2J) where 20 and 21 $\mu l/g./hour$ was recorded for CO_2 evolution and O_2 uptake respectively in the zone close to the roots and 3·6 and 1·9 $\mu l./g./hour$ in the non-rhizosphere zone (soil 2A). One would expect a considerable part of this difference to have been due to the fact that the material surrounding the roots was high in organic matter, whereas that farther away was far more mineral in nature. An examination of the table will show much smaller but consistent differences in the *Dicranoweisia antarctica* (soils 1A and 1D) and *Poa foliosa* (soils 3A and 24) environments, both of which are relatively simple from an ecological point

of view. The first represents a borderline plant community just able to maintain itself under extremely unfavourable conditions of wind, cold, and poor light, and the second represents a community able to colonize beach sand directly adjacent to the intertidal zone. It would appear, then, that the rhizosphere influence becomes effective and important, at least under these conditions, with the earliest seral communities in a plant succession.

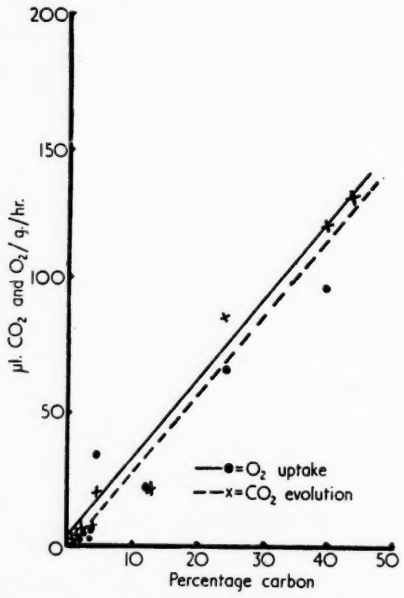


FIG. 1. The correlation between O_2 uptake and CO_2 evolution and the carbon content of Macquarie Island soils. (O_2) $y = 5{\cdot}9x+7$; (CO_2) $y = 5{\cdot}8x+0{\cdot}3$.

Further, it will be noted that the activity shown by soil 32, an organo-mineral rubble collected about the roots of the primary rock colonizing grass (*Puccinellia macquariensis*) is comparable with the activity of the highly organic and relatively well-developed soil 2J.

The correlation coefficients at 25° C. between (1) CO_2 evolution and (2) O_2 consumption and the carbon content of the soil were found to be $+0{\cdot}985$ and $+0{\cdot}965$ respectively. It has been found that there is a straight-line relationship between these two variables. This is represented in Fig. 1, in which will also be seen the close agreement between the curves for oxygen uptake and CO_2 evolution. The difference between the two curves is not significant, so that R.Q. does not vary as the carbon content of the soil increases. From results obtained by Rovira (unpublished) it would appear that individual variations in R.Q. (Table 5) are caused by such factors as local variations in moisture content, aeration, pH, and C/N ratios.

Discussion

Differences between the numbers of micro-organisms in the three non-rhizosphere samples are not related to the organic-matter content of the soils. Sample 4, which has the highest number is a beach sand, with good aeration, high mineral content, and higher mean temperatures compared with the moraine muds of samples 1 and 2, and these factors may be responsible for the higher counts. A further factor could be the contamination by seals and penguins and the decomposition of kelp.

The masses of soil and root in each sample have been included in Table 2 and results are expressed on two bases: (*a*) per gramme of soil in sample for both rhizosphere and non-rhizosphere; (*b*) per gramme of roots in sample for the rhizosphere. The results are expressed on the two bases because it has been shown (Clark, 1947, and Rovira, 1953) that the ratio of root mass to soil mass in samples can greatly influence the results when expressed on a soil-mass basis.

The expression of results per gramme of soil permits the detection of a stimulation of the micro-organisms in the soil surrounding the root but may not be suitable for comparing the rhizosphere effects of different root systems in the same soil and plants in different soils. The reason for this is that the nature of the root system and the soil will influence the amount of soil adhering to the roots. The expression of results on a root-mass basis in studying the rhizosphere effects of normal pasture species is more satisfactory for assessing the activity of the different species as it is less dependent on the amount of soil adhering to the roots. However, owing to the nature of the rhizosphere samples used in these experiments in which the roots ranged from 5 mm. diameter to fine, fibrous roots, the expression of results per gramme of root does not prove very satisfactory—*Cotula plumosa*, with its fibrous root system is favoured in comparison with *Pleurophyllum Hookeri* which has thick fleshy roots.

In general, the rhizosphere effect is less than that obtained with plants from temperate regions, this being probably due to the less vigorous growth by the plants on Macquarie Island.

The low ratio of rhizosphere count to soil count in the *Azorella selago* and *Dicranoweisia antarctica* associations can be explained in terms of the nature of the 'rhizosphere', which included buried vegetative parts, together with the roots. The low rhizosphere count to soil count (R:S) ratio for the *Cotula plumosa* society is probably due to the very low root-mass/soil-mass ratio in this sample, as the rhizosphere effect is very high when expressed on a root-mass basis. The *Pleurophyllum Hookeri* association shows the highest R:S ratio, but this is probably due to the high root-mass/soil-mass ratio.

The influence of low soil temperatures on the temperature requirements of the isolates is shown in the results in which 96 per cent. of the isolates have a temperature range of 10° to 25° C. and 4 per cent. can grow at 37° C. In comparison with these results it has been found that 50 per cent. of the isolates from sub-tropical soils have a temperature range of 25° to 37° C. and 50 per cent. have a lower range of 17° to 25° C.

The findings that the plants did not show a selective stimulation of

gram-negative proteolytic organisms are surprising in view of results obtained in rhizosphere studies of Australian pasture plants (Rovira, 1953) and those reported by Clark (1940) and Lochhead (1940). This apparent non-selective stimulation may be due to the rather weak rhizosphere effect exerted by the plants studied.

It is a little surprising that such a good relationship should be found between the total carbon content of the soils and the respiration. Although organic analyses of the soils have not been carried out, one would expect considerable qualitative differences to exist in the organic-matter fractions of the various soils because of the differences in the conditions governing their formation. If there are qualitative differences in the organic-matter fractions of the various soils, then it appears that the amount of readily decomposable material must be proportional to the total organic-matter content of the soil.

Although the respiration of these highly organic soils at 25°C. is high compared with that of a subtropical soil, it is not sufficient to keep pace with the deposition of plant litter. This must mean that the activity of the soil micro-organisms is limited by the low temperatures prevailing on the island, otherwise it should be expected that every fraction of the organic matter should readily be fully humified and that a less primitive soil would develop. The effect of temperature on respiration has been studied and is reported elsewhere (Bunt and Rovira, 1955).

Acknowledgements

We wish to thank the following: The Australian National Antarctic Research Expedition for allowing one of us (Bunt) to return to Macquarie Island during the 1953 changeover operations to collect the samples for these studies; the University of Sydney for providing laboratory facilities in the School of Agriculture; Mr. J. M. Vincent for his advice and criticism of the manuscript; and Mr. R. Eatock for his technical assistance.

REFERENCES

Bunt, J. S., and Rovira, A. D. 1955. The effect of temperature and heat treatment on soil metabolism. J. Soil Sci. **16,** 129.

Cheeseman, T. F. 1919. The vascular flora of Macquarie Island. Australasian Antarctic Expedition, 1911–14. Scientific Reports, Series C—Zoology and Botany, vii. 3.

Clark, F. E. 1940. Notes on types of bacteria association with plant roots. Trans. Kansas Acad. Sci. **43,** 75.

—— 1947. Rhizosphere microflora as affected by soil moisture changes. Soil Sci. Soc. Amer. Proc. **12,** 239.

Lochhead, A. G. 1940. Qualitative studies of soil microorganisms, III. Influence of plant growth on the character of the bacterial flora. Canad. J. Res. C. **18,** 42.

Mawson, D. 1943. Macquarie Island, its geography and geology. Australasian Antarctic Expedition, 1911–14. Scientific Reports, Series A, vol. v.

Rovira, A. D. 1953. Use of the Warburg apparatus in soil metabolism studies. Nature, **172,** 29.

—— 1953. Some quantitative and qualitative aspects of the rhizosphere. Proc. Aust. Soil Sci. Conf. Adelaide, June 1953.

Taylor, B. W. (in press). The flora, vegetation and soils of Macquarie Island. A.N.A.R.E. Reports, Ser. B, **2,** pt. 1.

Walkley, A. 1947. A critical examination of a rapid method for determining organic carbon in soils. Soil Sci. **63,** 251.

THE EFFECT OF TEMPERATURE AND HEAT TREATMENT ON SOIL METABOLISM

J. S. BUNT* AND A. D. ROVIRA†

(*Microbiology Laboratories, School of Agriculture, University of Sydney*)

Summary

The respiration of subantarctic and subtropical soils was found to increase as temperature increased from 10° to 37° C. The gas exchange decreased between 37° and 50° C., but increased with further increase in temperature.

The oxygen uptake of soil subjected to various heat treatments showed an increase with the severity of treatment. The ratio of CO_2 evolved to oxygen consumed of heat-sterilized soil was always below 0·4, indicating a different oxidation process from that of unsterilized soils in which the ratio approximates 1·0. The activity of heat-sterilized soil was not inhibited by NaCl, NaN_3, or $HgCl_2$, indicating that a non-enzymic oxygen uptake and CO_2 evolution were occurring.

Chemical oxidation also seems to occur in unsterilized soils at high temperatures, and is likely to be a factor in the decline in organic matter in dry, exposed soils.

Oxidative enzymes of 'dead' micro-organisms appear to be chiefly responsible for the immediate respiration on wetting an air-dry soil.

Introduction

THE effect of temperature on CO_2 evolution of soils was studied by Deherain and Demoussey (1896), who found that respiration increased slightly from 20° to 45° C. and sixfold from 45° to 65° C. due to the action of the thermophilic population. Destruction of organisms between 65° and 80° C. resulted in a decreased respiration, but above 80° C. there was rapid evolution of CO_2, presumably due to chemical oxidation. Jensen (1934) found that an increase from 15° to 28° C. doubled respiration. Raising the temperature further to 37° C. increased respiration by another 50 per cent. Both investigations were restricted to CO_2 evolution, which need not be fully indicative of the metabolism of the soil population.

This paper reports the effects of temperature and heat treatment on both the oxygen uptake and the CO_2 evolution of one subtropical and five subantarctic soils.

Methods

The Warburg respirometer method of Rovira (1953), in which 7 g. (wet weight) of soil is placed in each of four flasks, two of which have KOH in the centre cups, was used to obtain duplicate measurements of oxygen uptake and CO_2 evolution at temperatures ranging from 10° to 73·5° C. Moisture content was kept constant for each soil in all the experiments, as it has been shown (Rovira, 1953) that moisture content can influence the respiratory quotient of the soil. In the experiments in which the effect of heat treatment on respiration was studied, Lismore soil (subtropical) and one of the more active of the Macquarie Island (subantarctic) soils were used. The soils were heated in the Warburg

* Biologist, Australian National Antarctic Research Expedition.

† Present address: Institute of Microbiology, Royal Agricultural College, Uppsala, Sweden.

flasks and sterile water added to restore the moisture content to its original level.

Chemical inhibitors were added in solution to air-dry Lismore soil to ensure adequate penetration to the sites of microbial activity. As the Macquarie Island soil, which contains approximately 120 per cent. water (oven-dry basis), could not be air-dried without radical alteration of its structure and general properties, the chemical inhibitors were added in solid form to give the same concentration in the soil water as in the Lismore soil.

Modified soil-extract agar (Bunt and Rovira, 1955) was used for total plate counts.

Organic Matter Content of Soils

Macquarie Island: subantarctic

Soil 1 . . 3·6%
„ 3 . . 12·5%
„ 24 . . 1·3%
„ 31 . . 4·4%
„ 38 . . 43·6%

Lismore: subtropical 8·0%

Results

The results presented in Table 1 show the uptake of oxygen, evolution of CO_2, and the associated respiratory quotients for five Macquarie Island soils and Lismore soil, at temperatures from 10° to 56° C.

TABLE 1

Microlitres of Oxygen taken up and CO_2 evolved per gramme soil, and Respiratory Quotient after 2 hours

Soil no.	*1*			*3*			*24*			*31*			*38*			*Lismore*		
Temp. °C.	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*
10	1	−1	..	31	39	1·03	0	0	0	34	41	1·2	167	129	0·82	0	5	..
20	6	7	1·2	27	33	1·2	0·5	3	6·0	53	45	0·85	216	185	0·86	3	5	1·7
25	10	13	1·3	41	40	0·98	3	6	2·0	91	39	0·43	382	259	0·68	7	10	1·4
30	13	11	0·85	92	99	1·08	5	5	1·0	44	27	0·61	215	126	0·59	14	16	1·1
37	30	21	0·70	173	153	0·88	7	6	0·86	162	110	0·68	722	336	0·47	36	38	1·06
44	27	28	1·04	168	141	0·84	8	5	0·62	152	103	0·68	850	1,006	1·2	24	21	0·87
50	35	34	0·97	130	119	0·92	3	3	1·0	346	371	1·07	746	548	0·73	39	38	0·98
56	48	54	1·1	134	140	1·04	4	3	0·75	466	436	0·93	1,320	780	0·59	104	92	0·89

Bunt and Rovira (1955) have reported correlations of +0·965 and +0·985 between organic-matter content of soil and oxygen uptake and CO_2 evolution respectively. This relationship applies to the soils studied in this experiment with the exception of soil 31 which has a high respiration for its low organic-matter content. This high activity is probably due to the organic matter being derived mainly from readily decomposable penguin faeces (the sample being taken in the vicinity of a penguin rookery).

With the respiration proportional to the organic-matter content of the soil, it would be expected that the expression of the results per unit of organic matter would show the temperature at which the soil population

was respiring most efficiently. The results reported in Fig. 1, expressing respiration per gramme of organic matter, fail to show consistent lower optimum temperatures for respiration of Macquarie Island soils.

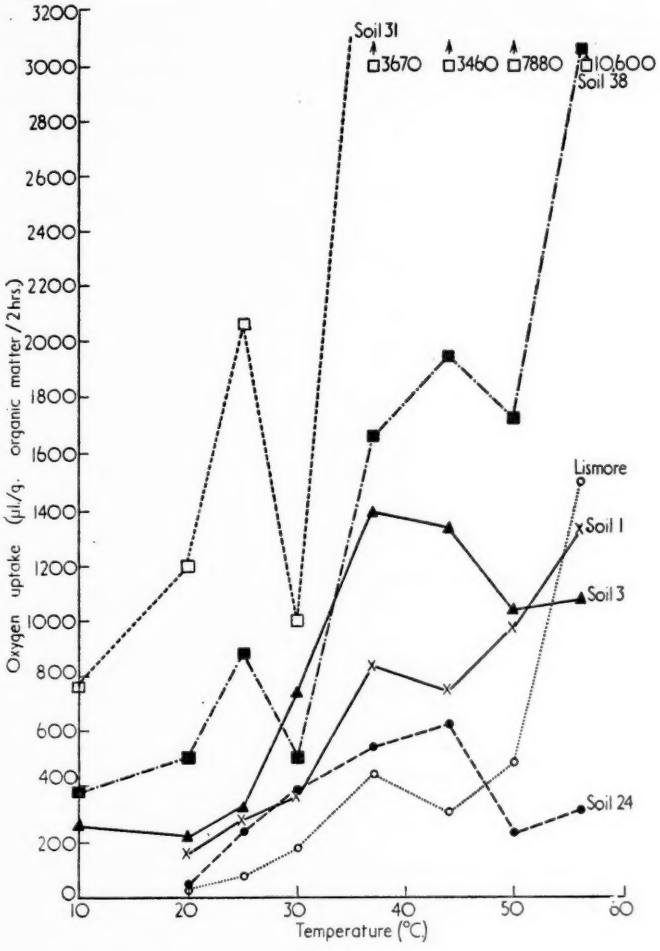


FIG. 1

A feature of the results in Fig. 1 is that the respiration for each soil reaches at least one peak at temperatures below 44° C. beyond which there is a decline in respiration followed by a further increase as the temperature is increased. In three Macquarie Island soils two respiration peaks occurred, one between 10° and 25° C. and a second at 37° or 44° C. These fluctuations in respiration below 44° C. could be due to

differential responses to increased temperature by the various groups of soil micro-organisms. However, it is difficult to explain the increased respiration above 50° C. in terms of microbial activity. As the temperature treatments were only of 2–4 hours duration, no development of a high thermophilic population could have occurred to account for the respiration above 50° C. Plate counts on Lismore soil during one respiration trial are reported in Table 2 and show a marked decrease after the soils have been exposed for 2 hours to a temperature of 44° C. or higher. Similar results were obtained with Macquarie Island soils.

TABLE 2

Plate Counts ($\times 10^6$) of Bacteria per gramme of Oven-dry Lismore Soil after 2 hours at each Temperature*

Temperature	*Plate count*
25° C.	0·205
30° C.	0·195
37° C.	0·135
44° C.	0·080
52·3° C.	0·025

* The initial count of the soil was $0{\cdot}133 \times 10^6$.

It seems reasonable to assume that the plate count is indicative of the resistance of the total soil population to high temperatures and that the high respiration in the presence of a much reduced population is due to one or more of three factors:

(*a*) an increased activity of surviving organisms;
(*b*) the activity of heat-stable enzymes;
(*c*) the chemical oxidation and decarboxylation of organic matter.

An attempt has been made to assess the importance of each of these factors by investigating the respiration of soils subjected to various heat treatments with and without chemical inhibitors (Tables 3–7).

TABLE 3

The Effect of Heat Treatment on Gas Exchange of Lismore Red Loam at 30° C. (μl./g. oven-dry soil/hr.)

	Untreated soil held at field capacity for 24 hrs.	*Heated to 85° C. for 2 days.*	*Steamed for 1 hr. on 3 successive days.*	*Autoclaved at 15 lb.*		
				1 hr.	*2 hrs.*	*3 hrs.*
O_2	19	1·4	5·9	7·0	8·5	9·0
CO_2	18	0·7	1·7	1·2	1·3	1·2
R.Q.	0·9	0·5	0·29	0·17	0·16	0·14

In Table 3 it can be seen that the maximum reduction of gas exchange was obtained when the soil was subjected to the lowest sterilization temperature (85° C. for 2 days). The oxygen uptake increased with the severity of sterilization while the CO_2 evolution was reduced to an almost constant level by all treatments.

Plate counts showed complete sterilization of the soils with each heat treatment, indicating that the oxidation is not due to the activity of viable cells. The ratio of CO_2 evolved to oxygen absorbed of approximately 1·0 in the unsterilized soil compared with 0·1–0·3 in heat-sterilized soils indicates that the oxidation reactions differ. The oxygen uptake by autoclaved soil could have been due to re-oxidation of groups reduced during heating while the CO_2 evolution must have been due to either a chemical decarboxylation or the action of heat-stable decarboxylases.

The contribution which the oxidation of an autoclaved soil may make to the decomposition of organic matter in unautoclaved soil at temperatures ranging from 25° to 73·5° C. was investigated with Lismore red loam. A soil which had been air dry for several months and which had been brought to field capacity immediately before commencing respiration measurements was included in the experiment to determine whether the high respiration rate obtained immediately upon wetting the soil (Rovira, 1953) was due to microbial activity, residual enzymes from organisms killed by desiccation, or a chemical oxidation.

TABLE 4

The Effect of Temperature on the Respiration of Unheated and Autoclaved Lismore Red Loams ($\mu l./g./hr.$)

Temperature °C.	*Soil held at field capacity for 24 hrs.*			*Soil freshly wetted to field capacity*			*Soil held at field capacity for 24 hrs. and autoclaved (15 lb. for 1 hr.) before experiment*		
	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*	O_2	CO_2	*R.Q.*
25	12	10	0·8	3·2	3·5	1·1	8	1·0	0·1
30	17	14	0·8	8	8	1·0	7	1·2	0·2
37	26	23	0·9	10	10	1·0	10	3·0	0·3
44	20	20	1·0	9	9	1·0	9	2·6	0·3
52·3	6·5	9	1·4	10	11	1·1	18	4·6	0·3
60	7	9	1·3	8	12	1·5	28	11	0·4
67	9	13	1·4	6	12	2·0	26	10	0·4
73·5	12	17	1·4	9	14	1·6	43	28	0·6

The respiration and respiratory quotients of the two unautoclaved soils responded similarly to temperature increases, which indicates that the immediate respiration upon wetting an air-dry soil is of a biological nature. Table 4 shows that in the three soils the ratio CO_2 evolved to oxygen absorbed increases as temperature increases. This indicates that the decarboxylation process, which occurs in the autoclaved soils and which proportionally increases more than oxygen uptake with temperature increase, may contribute significantly to the CO_2 evolution of unsterilized soil at the higher temperatures.

A study was made of the immediate respiration response to the addition of glucose to autoclaved and unheated soils to check whether autoclaved soils are capable of oxidizing glucose. The results reported in

Table 5 show that, at the temperatures indicated, the respiration of autoclaved soil showed no response to glucose supplement.

TABLE 5

The Effect of Glucose on the Respiration of Unheated and Autoclaved Soils (μl./g./hr.)

Temperature °C.	Unheated soil				Autoclaved soil			
	No glucose		Glucose		No glucose		Glucose	
	O_2	CO_2	O_2	CO_2	O_2	CO_2	O_2	CO_2
	(A) Lismore soil							
25	12	10	21	19	8	1	7·5	2·3
30	17	14	35	35	7	1·2	7·0	1·5
44	26	23	61	63	9	2·6	8·0	2·3
52	10	9	15	17	10	4·3	10	4·2
60	23	19	15	21	12	3·5	13	2·8
	(B) Macquarie Island soil							
25	22	21	49	42	12	2·8	14	5·7
30	30	25	62	57	14	3·3	15	5·7
44	40	40	56	51	19	6·5	20	9·0
52	33	30	38	33	27	13	30	16
60	26	28	28	29	32	17	35	22

In the unautoclaved soils glucose increased the respiration until 52° C. was reached, above which there was no difference due to glucose, which indicates that the respiration above 52° C. was due to the oxidation of other substances present in the soil.

If the formation of unstable oxidizable groups during autoclaving is a factor responsible for the oxygen uptake (as was suggested in the discussion of Table 3) then the oxidation rate would decline sharply with time after heating. Table 6 shows the respiration rates at various times following autoclaving and steaming.

TABLE 6

Oxygen Uptake and CO_2 Evolution at 30° C. of Heated Lismore Soils at Various Periods after Heating (μl/g./hr.)

Time after treatment	Autoclaved 15 lb. for 1 hr.			Steamed for 1 hr. on 3 successive days		
	O_2	CO_2	R.Q.	O_2	CO_2	R.Q.
1 hour	25	8·6	0·3	5·9	1·7	0·3
5 hours	13	4·3	0·3	3·6	1·2	0·3
6·5 hours	12	3·9	0·3	3·2	1·0	0·3
9 days	0·85	0·6	0·7	..	..	..

The results clearly indicate that the high respiration after autoclaving is only temporary and hence probably largely due to the formation of reduced substance or the exposure of readily oxidizable groups due to heating. However, the low rate of respiration with a R.Q. of 0·7 after

9 days is not likely to be due to autoclaving, and may be the level of non-microbial oxidation occurring in unheated soils.

In an attempt to show conclusively that the respiration of unheated soil at 30° C. is due to microbial and enzymic action, and that the activity of freshly autoclaved soil is non-enzymic, a further experiment was conducted to study the effect of three biological inhibitors on unheated and autoclaved soils. The results are given in Table 7.

TABLE 7

Oxygen Uptake and CO_2 Evolution of Unheated and Autoclaved Soils in the Presence of Biological Inhibitors at 30° C. (μl./g./hr.)

	Lismore soil			*Macquarie Island soil*		
	No inhibitor	NaN_3	NaCl	$HgCl_2$	*No inhibitor*	NaCl
	(A) Unautoclaved soil					
O_2	18	2·0	0·9	3·0	30	0
CO_2	17	3·0	1·3	2·5	25	4·2
R.Q.	0·9	1·5	1·4	0·8	0·8	..
	(B) Autoclaved soil					
O_2	18	15	21	20	14*	39
CO_2	2·9	0·7	1·7	1·9	3·3	8
R.Q.	0·2	0·05	0·1	0·1	0·2	0·2

* Readings taken 5 hours after autoclaving, hence the discrepancy between these results and those in final column (see Table 6: temporary effect of autoclaving).

In the unheated soils, the chemicals all exerted a strong inhibitive action with NaCl as the most effective inhibitor. The low respiration which occurred in the presence of the various inhibitors was probably due to incomplete penetration of the inhibitors into the soil interstices, allowing some microbial respiration to continue, and to some chemical oxidation of the type suggested in the discussion of the activity occurring 9 days after autoclaving (Table 6). The R.Q. of approximately 1·0 indicates that this residual respiration is probably mainly microbial. The failure of the biological inhibitors to inhibit gas exchange in freshly autoclaved soils indicates that this oxidation is non-enzymic.

Discussion

In the discussion of the results in Table 1 and 2, three factors were listed as possibly being responsible for the respiration of soils at temperatures above 44° C. Of these, it appears that the two important factors are the activity of heat-stable enzymes and chemical oxidation.

Oliver (1952)* has reported that decarboxylases of *E. coli* and *Strept. diacetilactis* remain active after the cells are killed by exposure to 80° C. for 5 min. or momentary exposure to 100° C, which supports the hypothesis of heat-stable enzymes in the soil. His finding that the heat stability

* The authors wish to draw attention to the later publication of R. H. Robertson and W. H. Oliver, J. Gen. Microbiol. **11**, 130, reporting that the results quoted above are not valid.

of these enzymes is increased in complex media indicates that in the soil, where the enzymes may be protected by the soil organic matter, the heat stability will be increased.

It has been shown that the respiration of freshly wetted soil and soil which had been at field capacity for 24 hours responds similarly to increases in temperature. The similarity in response indicates similar oxidation reactions, although from the plate counts (Rovira, unpublished data, 1953) the micro-organisms must be respiring more actively in freshly wetted soil, or some activity must be due to residual enzymes associated with organisms incapable of multiplication on the medium provided in the plate count.

The hypothesis that enzymes can exist in soil in cells incapable of multiplication in the medium used in the plate count provides an explanation for the low correlation between plate count and respiration and the high correlation between total count and respiration found by Jensen (1936) and Vandecaveye and Katznelson (1938) and also for the high respiration immediately an air-dry soil is wetted (Rovira, 1953).

The chemical oxidation which probably occurs at a very slow rate at normal temperatures could be important in sparsely vegetated areas where, during the dry season, high surface soil temperatures occur. The decline in the organic-matter content of inland soils which are too dry for most of the year for microbial and enzymic decomposition, could be due to chemical oxidation, since soil temperatures of 47°–50° C. for several hours per day occur even in cleared subtropical coastal soils (Jenkins, personal communication).

Acknowledgements

The authors wish to thank Mr. J. M. Vincent, Senior Lecturer in Agricultural Microbiology, University of Sydney, for helpful criticism and suggestions, and Mr. R. Eatock for technical assistance throughout the investigations.

Sampling of soils was carried out by the first author whilst a member of the 1951–2 Australian National Antarctic Research Expedition to Macquarie Island.

REFERENCES

Bunt, J. S., and Rovira, A. D. 1955. Microbiological studies of some subantarctic soils. J. Soil Sci. **6,** 119.

Deherain, P. P., et Demoussey. 1896. Sur l'oxydation de la matière organique du sol. Ann. Agron. **22,** 305–37.

Jensen, H. L. 1936. Contributions to the microbiology of Australian soils. Proc. Linn. Soc. N.S.W. **61,** 27–55.

Oliver, W. H. 1952. The stability of some bacterial enzymes toward heat and chemical bactericides. J. Gen. Microbiol. **7,** 329–32.

Rovira, A. D. 1953. Use of the Warburg apparatus in soil metabolism studies. Nature, **172,** 29.

Vandecaveye, S. C., and Katznelson, H. 1938. Microbial activities in soil. Soil Sci. **46,** 139–67.

MEASUREMENTS OF THE ACTIVITIES OF BASES IN SOILS

R. K. SCHOFIELD AND A. WORMALD TAYLOR

(Rothamsted Experimental Station, Harpenden)

In 1930 Teräsvuori argued, on the basis of the Donnan membrane equilibrium, that there should be a constant value of the ratio $\sqrt{([Ca])}/[H]$ for all the solutions in equilibrium with a given soil sample which contains calcium as the dominant exchangeable cation. He obtained a rough experimental verification of this conclusion by a study of the pH values of soil samples suspended in water and calcium chloride solutions of concentrations up to 5×10^{-3} M.

Schofield (1947) also checked this relation in the case of a surface soil taken from the Rothamsted Park Grass plots, but found that it did not hold for a sample of a red subsoil from Natal. It was known from previous work that, whereas the Rothamsted soil was essentially a negatively charged colloid, the Natal soils carried approximately equal amounts of positive and negative charges. He therefore concluded, in agreement with Mattson and Wiklander (1940), that the ratio of $\sqrt{([Ca])}/[H]$ may only be expected to remain constant when the soil carries negative charges alone, or an overwhelming proportion of negative charge.

In all these earlier experiments dilute solutions were used, and no corrections for activity coefficients were applied, but in the experiments described in the present paper, where the constancy of the ratio $\sqrt{(a_{Ca})}/a_H$ is tested over a wider range of concentration, this correction becomes important, and activities have been substituted for concentrations throughout.

Precise meaning can be given to the ratio $\sqrt{(a_{Ca})}/a_H$ by multiplying both the top and bottom by the activity of any anion present, Cl^- or OH^-:

$$\frac{\sqrt{(a_{Ca})}}{a_H} = \frac{\sqrt{(a_{Ca})} \times a_{Cl}}{a_H \times a_{Cl}} = \frac{\sqrt{(a_{Ca})} \times a_{OH}}{a_H \times a_{OH}} \qquad (1)$$

when it is seen that, since the product $a_H \times a_{OH}$ is constant, the ratio $\sqrt{(a_{Ca})}/a_H$ is a direct function of the activity of calcium hydroxide in the system.

In the experiments described in this paper the ratio is measured in the form $\sqrt{(a_{CaCl_2})}/a_{HCl}$ since these activities may be easily measured and, as they are both electrically neutral substances, their activities may be expressed without any of the ambiguity associated with the activities of individual ions.

Equation (1) may also be written:

$$pH - \tfrac{1}{2}pCa = p(HCl) - \tfrac{1}{2}p(CaCl_2) = \tfrac{1}{2}\log a_{Ca(OH)_2} + 14{\cdot}2, \qquad (2)$$

where pH, pCa, p(HCl), and $p(CaCl_2)$ refer respectively to the negative logarithms of the activities of hydrogen and calcium ions, and hydrochloric acid and calcium chloride in the equilibrium solution; and 14·2 is the corresponding value for the ion product of water at 20° C.

Using the equation in this form we obtain a measure of the chemical potential of calcium hydroxide in the system, since this is thermodynamically defined by the function:

$$\mu = \mu_0 + \frac{RT}{2{\cdot}303} \log a_{Ca(OH)_2}, \qquad (3)$$

where μ is the chemical potential and μ_0 the chemical potential in the standard state.

Experimental Method

The design of the experiments fell into two stages:

1. The preparation of the soil sample. 400 g. of a sample of surface soil from Sawyers 1 field on the Rothamsted farm were leached in a Büchner funnel with 2 litres of M $CaCl_2$ solution followed by 8 litres of M/10 $CaCl_2$ containing N/4,000 HCl. The soil was then washed as free as possible of the solution with water under suction and allowed to drain for 24 hours.

Sawyers 1 field, from which the soil was taken, is an arable field which has not received any special experimental treatment for a number of years: the soil is non-calcareous with a pH value 4·60 in M/100 $CaCl_2$ and contains about 20 per cent. clay, which has been found by X-ray diffraction to consist of about 40 per cent. illite, 40 per cent. kaolin, and 20 per cent. vermiculite. The soil also contains about 2 per cent. of organic matter and has a total base exchange capacity of 10 m.e. per 100 g.

2. The determination of the activities of calcium chloride and hydrochloric acid in a series of solutions in equilibrium with the prepared soil. The bulk sample was split into 8 sub-samples of 50 g. each, each one being shaken up with 100 ml. of a calcium chloride solution, the concentrations of which ranged from 10^{-1} to 10^{-4} M.

The value of a_{HCl} was measured electrometrically in each suspension, and after filtration the values of the calcium and chloride ion concentration were determined in the filtrate. In the sub-samples shaken up with solutions more dilute than about 3×10^{-3} M, a certain amount of deflocculation of the clay occurred and clear filtrate could only be obtained by filtration through a double Whatman No. 50 filter paper.

Measurement of a_{HCl}

The value of the activity of the hydrochloric acid in the suspension was determined by measurement of the e.m.f. from a cell consisting of a silver/silver chloride electrode, made by electrolytic deposition of silver chloride on a clean silver wire, and a glass electrode supplied by Marconi Instruments Ltd. The cell was calibrated by measurement of the e.m.f. obtained from a series of accurately made-up HCl solutions of known activity. The e.m.f. values were measured with a Marconi Type 717A pH meter giving readings accurate to ± 1 millivolt.

The measurements on the soil suspensions were made by placing the electrodes in the soil suspensions immediately after shaking. Steady and reproducible readings were always obtained as soon as the sand fraction

of the soil had settled and turbulence had ceased. Further sedimentation did not produce any variation, even when the whole soil had settled and the electrodes were surrounded by the clear equilibrium liquid.

In order to avoid any error arising from temperature differences between the solutions, the calibration measurements using hydrochloric acid were carried out at the same time and check calibrations were made during the soil measurements.

It should be noted that the cell does not contain any liquid junction, and the determination of the hydrochloric acid activity by this means is, therefore, not subject to the criticisms which may be made if the pH value were determined using a saturated calomel electrode with a saturated KCl junction tube. In such a system the total e.m.f. of the cell might be affected by variations in the liquid-junction potential owing to the presence of dispersed colloidal particles in the system. Where the electrolyte in the solution is dilute and the potential difference across the double layer around the particles is large, the error due to this cause may become quite serious and reproducible results difficult to obtain.

Calculation of a_{CaCl_2}

The calcium ion concentration in the filtered equilibrium solution was measured spectrographically and the chloride ion by titration using the differential electrometric titration method of MacInnes and Dole. This method allows the amount of chloride in 10 ml. of an N/1,000 solution to be determined to the nearest 0·1 m.e. per litre (or about ±5 per cent.), but the percentage accuracy is naturally much improved at higher concentrations.

When the total analyses of the solution had been completed, it was found that the chloride ion concentration was greater than the sum of the calcium and hydrogen. The proportion of this excess chloride increased with dilution, amounting to about 25 per cent. in the most dilute ($[Cl] = 1{\cdot}9 \times 10^{-3}$ M). Further spectrographic examination showed that the solutions contained appreciable quantities of potassium ion in addition to the calcium. This is probably due to the release of this ion from the illitic clay mineral present under the acid conditions during the preparation of the bulk soil sample.

In the calculation of the ionic strength of the solutions it was therefore assumed that they contained potassium chloride in addition to the calcium chloride at a concentration equal to that of the excess chloride ion.

The activity coefficient of calcium chloride ($\gamma^{\pm}_{CaCl_2}$) was calculated from the ionic strength of the solution (μ) using the Debye–Hückel relationship

$$-\log \gamma^{\pm}_{CaCl_2} = \frac{\sqrt{\mu}}{1+1{\cdot}6\sqrt{\mu}} - 0{\cdot}15\, m, \qquad (4)$$

where m is the molarity of the $CaCl_2$. The value of $p(CaCl_2)$ is then given by the relation

$$p(CaCl_2) = -\log[Ca] + (-2\log[Cl]) + 3(-\log \gamma^{\pm}).$$

The complete experimental data are given in Table 1, and the final

values of p(HCl) and $\frac{1}{2}$p($CaCl_2$) obtained for the equilibrium solution are plotted in Fig. 1.

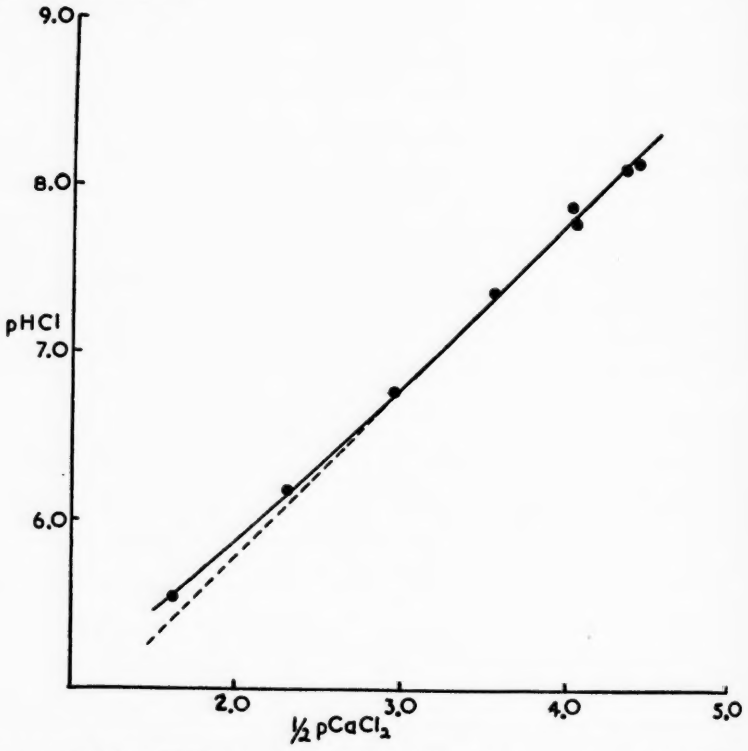


FIG. 1. Activity of HCl found after equilibriating a soil sample with solutions of different $CaCl_2$ activities. The equation for the straight line is

$$p(HCl) - \tfrac{1}{2}p(CaCl_2) = 3{\cdot}78.$$

It will be seen that the points obtained lie close to a straight line of slope 1·00 except for the most concentrated solution (above 3×10^{-2} M) where there is appreciable divergence from the straight-line relationship.

TABLE 1

p(HCl)	Ca *molar*	Cl *molar*	μ	$\log \gamma^{\pm}_{CaCl_2}$	$\frac{1}{2}$p($CaCl_2$)
5·54	$9{\cdot}4 \times 10^{-2}$	$2{\cdot}02 \times 10^{-1}$	0·296	0·277	1·62
6·18	$2{\cdot}95 \times 10^{-2}$	$5{\cdot}68 \times 10^{-2}$	0·088	0·197	2·31
6·76	$8{\cdot}5 \times 10^{-3}$	$1{\cdot}90 \times 10^{-2}$	0·0275	0·131	2·95
7·88	$1{\cdot}5 \times 10^{-3}$	$3{\cdot}12 \times 10^{-3}$	0·0046	0·061	4·01
7·36	$3{\cdot}14 \times 10^{-3}$	$6{\cdot}80 \times 10^{-3}$	0·0100	0·086	3·55
8·10	$7{\cdot}7 \times 10^{-4}$	$1{\cdot}95 \times 10^{-3}$	0·0028	0·049	4·34
8·14	$6{\cdot}8 \times 10^{-4}$	$1{\cdot}75 \times 10^{-3}$	0·0024	0·045	4·41
7·78	$1{\cdot}29 \times 10^{-3}$	$3{\cdot}10 \times 10^{-3}$	0·0044	0·060	4·04

Aluminium hydroxide potential

In a soil where aluminium is the predominating exchangeable cation, the ratio controlling the equilibrium concentration of the ions is $\sqrt[3]{(a_{Al})}/a_H$ and the corresponding form of equation (2)

$$pH-\tfrac{1}{3}pAl = p(HCl)-\tfrac{1}{3}p(AlCl_3) = \tfrac{1}{3}\log a_{Al(OH)_3}+14{\cdot}2. \qquad (5)$$

The experimental method was unchanged: the soil was prepared by leaching with 5 litres of M/10 $AlCl_3$ solution containing M/1,000 HCl, and then washed and drained as before.

In the final equilibrium suspensions the aluminium chloride solutions varied from 1×10^{-2} to 5×10^{-5} M; the aluminium ion concentrations in the equilibrium filtrates were determined colorimetrically by Chenery's method (1948) using aluminon reagent. The values for a_{HCl} and the chloride ion concentration were found as before.

It was again found that the chloride ion concentration was in excess of the sum of the aluminium and hydrogen, owing to the presence of potassium in the solutions. The proportion of excess chloride was, however, greater than in the calcium solutions, amounting to 70 per cent. of the total in the most dilute ($[Cl] = 1{\cdot}3\times10^{-3}$ M).

The ionic strengths of the equilibrium solutions were calculated as before, making allowance for the potassium present. The activity coefficient of the aluminium chloride was then obtained from the relation

$$-\log\gamma^{\pm}_{AlCl_3} = \frac{1{\cdot}5\,\mu}{1+2{\cdot}6\,\mu}-1{\cdot}4\,m. \qquad (6)$$

The experimental results are plotted in Fig. 2: they are similar to those obtained in the calcium experiment, except that departure from the linear relation takes place at a much lower salt concentration.

Sodium and potassium hydroxide potentials

Here the constant ratios are a_{Na}/a_H and a_K/a_H and the appropriate equations

$$pH-pNa = p(HCl)-p(NaCl) = \log a_{NaOH}+14{\cdot}2, \qquad (7)$$

$$pH-pK = p(HCl)-p(KCl) = \log a_{KOH}+14{\cdot}2, \qquad (8)$$

but in these cases the experimental technique was modified owing to difficulties caused by the higher concentrations at which deflocculation of the soil took place, preventing satisfactory filtration of the suspensions.

The sodium soil was prepared by washing with about 5 litres of M NaCl at pH 3·0 and the potassium with 10 litres of M/10 KCl at pH 4·0. In the potassium soil the concentration of KCl was reduced to below 10^{-3} by washing on the pump before the soil became impermeable, but no washing at all was possible with the sodium soil. The entrapped liquid was, therefore, removed by shaking up the soil with water and separating it in a Sharples super-centrifuge. This process was repeated until the electrolyte concentration had fallen to below M/1,000.

On shaking up the sub-samples with the electrolyte, deflocculation occurred in all the potassium suspensions below about 0·2 M and in the

sodium below 0·1 M, making filtration impossible. Where this occurred, the solutions were flocculated by the addition of 4 ml. of 2N ammonium nitrate solution after the hydrochloric acid activity had been measured as before. The suspensions were then filtered and measured in the normal way.

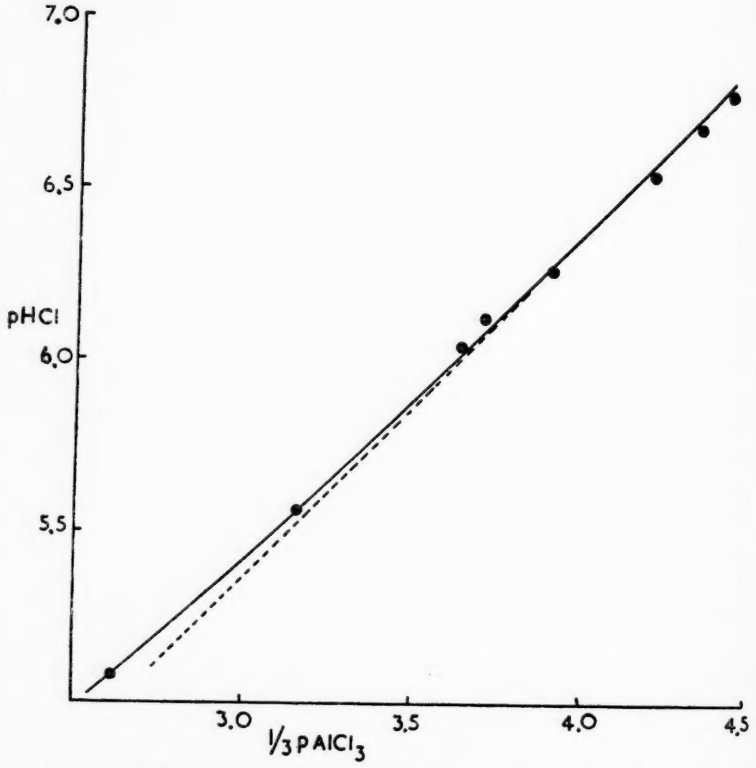


FIG. 2. Activity of HCl found after equilibriating a soil sample with solutions of different $AlCl_3$ activities. The equation for the straight line is

$$p(HCl) - \tfrac{1}{3}p(AlCl_3) = 2{\cdot}36.$$

The cation concentration could no longer be determined directly since the addition of the ammonium ion had upset the cationic equilibria in the suspensions. The filtered solutions were therefore examined spectrographically for calcium and magnesium as being the most likely ions present in addition to the sodium or potassium. Their concentrations were found to be below 1×10^{-4} even in the presence of the 0·16 M ammonium, and it was therefore considered that their concentrations could be neglected in the original suspensions.

Since, also, the ratio of [K]/[H] and [Na]/[H] is about 1,000, the values

of the cation concentrations were calculated assuming that the hydrogen ion could be neglected and that

$$[K] = [Cl] \quad \text{or} \quad [Na] = [Cl].$$

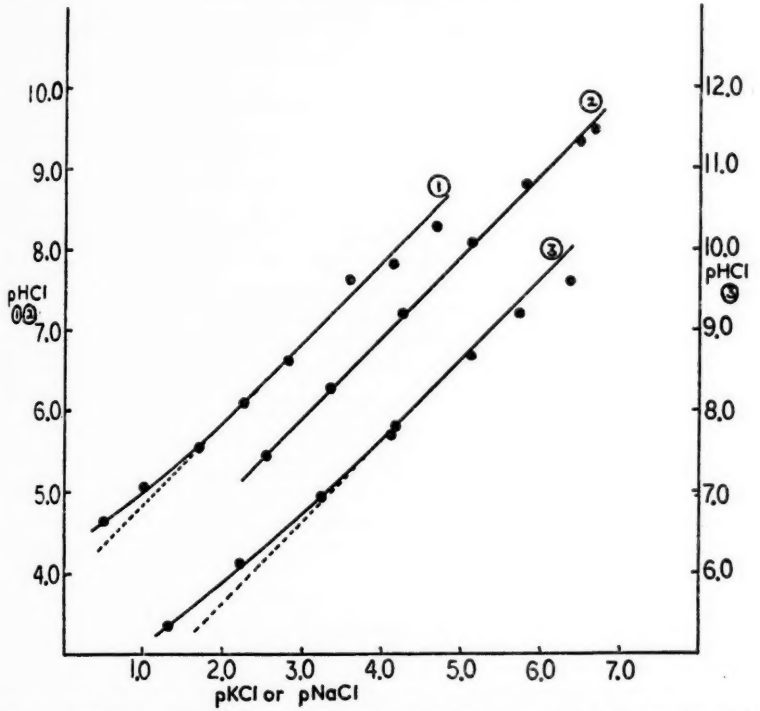


FIG. 3. Activities of HCl found after equilibriating appropriate soil samples with solutions of different KCl or NaCl activities. The equations for the straight lines are:

(1) $p(HCl)-p(KCl) = 3{\cdot}80$; (2) $p(HCl)-p(KCl) = 2{\cdot}90$;
(3) $p(HCl)-p(NaCl) = 3{\cdot}62$.

The error involved in determining the cation concentration is quite small, but in the more dilute solutions the presence of a large quantity of ammonium nitrate introduces a systematic error which will be discussed below.

The activity coefficients of the sodium and potassium chlorides were calculated from the relations

$$-\log \gamma^{\pm}_{NaCl} = \frac{0{\cdot}5\,\mu}{1+1{\cdot}4\,\mu} - 0{\cdot}14\,m, \tag{9}$$

$$-\log \gamma^{\pm}_{KCl} = \frac{0{\cdot}5\,\mu}{1+1{\cdot}3\,\mu} - 0{\cdot}047\,m. \tag{10}$$

Two experiments were carried out with different potassium soils, and one with a sodium. The results of all of these are plotted in Fig. 3.

Negative adsorption

If the difference of electrical potential across the electrical double layer surrounding a charged colloidal particle is large, as it will be when the concentration of electrolyte in the solution in which the particle is suspended is low, then the ions carrying the same charge as the particle will be strongly repelled from the surface and the concentration in the external solution will be increased.

This effect of the negative adsorption of chloride ions by soil clays has previously been examined by Schofield (1947) and by Mattson (1929), working on clays suspended in solutions of sodium salts.

If such a suspension is filtered, the soil residue being allowed to drain naturally without any attempt being made to express the entrapped solution, the filtrate obtained represents fairly closely the composition of the equilibrium solution and a true value for the equilibrium chloride concentration will be obtained by analysis of the filtrate.

In the sodium and potassium experiments recorded above this is not the case. Where the ammonium nitrate has been added and the total electrolyte concentration brought up to about 0·2 M, the potential difference has been reduced so that flocculation occurs and the chloride ion is no longer negatively adsorbed. Consequently, the chloride ion concentration in the filtrate is less than for the solution in equilibrium with the original suspension where the value of a_{HCl} was measured, and the calculated value of p(NaCl) is higher than in the original system.

On the potassium curves the effect does not appear to be large, but in the sodium curve the displacement of the points to the right can clearly be seen.

Discussion

It will be seen from Figs. 1, 2, and 3 that the same type of curve is obtained in all four cases. For a large portion of the curve the points obtained lie about a straight line of slope 1·00, but where the chloride concentration has risen above a certain level—different in each case—the experimental points lie above this line. The linear portion of the curve indicates that up to a limiting chloride concentration there is a characteristic value of the base activity associated with the soil sample which is independent of the chloride concentration. If, however, this is increased and rises above the limiting concentration, the base activity depends increasingly on the chloride concentration and is then no longer characteristic of the soil sample alone.

This deviation from the straight-line relationship is much greater than may be accounted for by the experimental error in the measurements, which does not exceed $\pm 0{\cdot}02$ to $\pm 0{\cdot}03$ units on the logarithmic scale employed in the diagrams or by any error in the calculated activity coefficients of the salts which is of the same order of magnitude. The experimental error is increased in the more dilute solutions owing to analytical difficulties, but except in the sodium experiment—discussed above—the points are scattered about the straight line and show no drift.

The effect of the higher electrolyte concentration on the base activities may be explained by considering the distribution of ions in the electrical double layer. When the electrolyte concentration of the system is low, there is a large difference of electrical potential between the region of solution close to the negatively charged surface and the equilibrium solution at a great distance from it; associated with this there is a large gradient of ionic concentration, the cations being attracted to the surface and the anions repelled. When the potential difference is sufficiently large for the chloride ion concentration close to the surface to be very small, the composition of the solution close to the surface is scarcely affected by changes in the chloride concentration in the outer solution. In these circumstances the activities of the bases remain practically constant.

If, however, the chloride concentration is sufficiently increased, the potential difference falls and eventually the chloride ions will no longer be excluded from the surface region of the solution and the surface conditions cease to be independent of the outer concentration. When this happens the value of the hydroxide activity changes as the surface conditions are affected, and the experimental points deviate from the straight line of slope 1·00.

The approximate concentrations at which this occurs for the four ions examined are Ca −N/25, Al −N/200, K −N/10, and Na −N/25. It should be emphasized that these figures only refer to the chloride salts; these were chosen for the experimental investigation because it is believed that the chloride ion is the one causing least disturbance to surface conditions. An anion such as sulphate, showing general positive absorption, causes disturbance of the surface conditions at a much lower level of electrolyte concentration.

Effect of soil type

It may also be shown from the Gouy theory that the base activity of a soil is not so readily measured in soils where the density of negative charge on the soil particles is low. When this is the case, the accumulation of cations in the surface layers of solution is less, and the controlling activity of the hydroxide is more sensitive to changes in the outer electrolyte concentration; it may then only be measured using very dilute solutions.

In the previous experiments described by Schofield the ratio law was found to be inapplicable to a red clay loam from Natal, a soil well known to have an isoelectric point of charge, probably on account of the large number of positive charges developed on the iron oxide present. A similar result was obtained by Mattson and Wiklander with a ferruginous laterite. The Rothamsted Grass Plot soil previously used by Schofield and the soil used in the present experiments are surface soils containing both organic matter and a fairly high content of illite clay.

Although in the experiments described in this paper the soils examined were specially prepared to contain only one other exchangeable cation in addition to the hydrogen, the base activity of any of the exchangeable cations present in a field soil may be determined in a similar way

provided, as in the original experiment described by Schofield, care is taken to ensure that the solution examined has been brought into a true equilibrium with the soil without disturbing the base status.

Summary

The activities of aluminium, calcium, potassium, and sodium hydroxides have been determined in suspensions of soil in chloride solutions of widely differing concentrations by measuring the ratio of activity of the metal chloride to the activity of hydrochloric acid. This procedure avoids the uncertainties inherent in attempts to measure the activities of individual ions.

For each of the four ions used there is a range of dilute chloride solutions within which the hydroxide activity is easily measurable and is nearly independent of the chloride concentration. The method is inapplicable in the absence of chloride, and experimental difficulties were encountered at the lowest concentrations used.

At higher salt concentrations there is a distinct rise in the hydroxide activity with increase in salt concentration. Both this rise and the substantial constancy at the lower chloride concentrations are to be expected for suspensions of negatively charged colloids exerting buffer action. The constant hydroxide activity approached as the chloride concentration is reduced is evidently a characteristic of the soil sample in question.

Acknowledgements

In addition to the Agricultural Research Council, for an Agricultural Research Grant awarded to Mr. Taylor during the tenure of which this work was carried out, the authors wish to express their thanks to Mr. H. H. Le Riche of the Soil Survey of England and Wales, who undertook the spectrographic analyses, and to Mrs. J. A. Parker of the Physics Department for her help in some of the preliminary experimental work.

REFERENCES

Chenery, E. M. 1948. The Analyst, **73,** 501.
MacInnes, D. A., and Dole, M. 1929. J. Amer. Chem. Soc., **51,** 1119.
Mattson, S. 1929. Soil Sci. **28,** 179.
—— and Wiklander, L. 1940. Ibid. **49,** 109 and 135.
Schofield, R. K. 1947. Proc. Eleventh Internat. Congr. Pure and App. Chem. **3,** 257.
—— 1947. Nature, **160,** 408.
Teräsvuori, A. 1930. Valtion Maatalonskoetorminnan Julkaisuja No. 29. Helsinki.

(*Received 4 December 1953*)

THE RATE OF WATER UPTAKE OF SOIL CRUMBS AT LOW SUCTIONS*

W. W. EMERSON

(*Rothamsted Experimental Station, Harpenden*)

I. LABORATORY INVESTIGATION

It was shown very clearly during the wet winter of 1951–2 that a serious structural problem in clay soils is that of maintaining the subsoil sufficiently permeable to prevent waterlogging. According to farmers, land drains better if it has carried a deep-rooted crop the previous summer. This suggested that once a clay subsoil has been dried out it might take an appreciable time to wet again completely.

The moisture in a clay soil is held partly by capillarity and partly as thick films between the negatively charged surfaces of clay crystals. When dry subsoil is slowly wetted, the capillary pores should refill as the suction is reduced to their characteristic water-entry value, provided the air to be displaced can escape. In practice, as the subsoil is slowly wetted from above over a period of weeks with succeeding falls of rain, the air will have ample opportunity to escape. It does not seem, therefore, that any delay in re-wetting can be attributed to the removal of capillary water by roots. There remains the possibility that after shrinkage of the clay, the water films might take a long time to regain their equilibrium thickness at low suctions. Schofield (1946) has given an exact expression for the relation between suction and the film thickness between two negatively charged plates immersed in an electrolyte where both the cations and anions are univalent. In soils, however, where calcium is the predominant cation, a corresponding formula is needed for an electrolyte containing divalent cations. The following figures show the variation of film thickness with suction for a calcium-chloride concentration of 1×10^{-2} M. calculated by a method developed by the author (Emerson, 1954).

Suction (dynes/cm.^{2}) .	10	1,000	10,000	100,000
Film thickness (mμ) .	11·9	9·93	7·90	3·84

A calcium-chloride concentration of 1×10^{-2} M. was chosen as being about the maximum concentration reached in the soil solution under English conditions.

It will be seen that a considerable rearrangement of the clay crystals within the crumbs will be necessary to accommodate the increasing thickness of film as the suction is reduced. (Note: 1 cm. of water suction equals 1,000 dynes/cm.^{2}) Also, at equilibrium, the repulsive pressure between the clay crystals must equal the attractive force between the

* This work forms part of a thesis approved for the Ph.D. degree of the University of London.

crystals, so that the cohesive strength of the crumbs will decrease correspondingly as the suction approaches zero.

Experiments were therefore made to find out how long soil crumbs would continue to absorb water when the base of the crumb is kept at zero suction.

Rate of Uptake of M./100 $CaCl_2$ by Soil Crumbs

Soils used. Soil samples were taken from the 0–4-in. layer of Barnfield 2 N—a plot on the Rothamsted permanent mangold field receiving 14 tons per acre f.y.m. together with a complete artificial fertilizer—and also from the 0–4-in. layer of adjoining permanent grass. The samples were taken when the soil was sufficiently wet to break down naturally into crumbs. In addition, short 3-in. diameter cores were taken from the Barnfield 8 O plot at a depth of about 18 in. The cores were afterwards cut to a convenient size in the laboratory with a pastry cutter. All the soil was dried to pF 4·2 (wilting-point) before use by leaving it for several weeks in a vacuum desiccator containing saturated barium nitrate.

Apparatus

The apparatus used is shown in Fig. 1. A crumb about $\frac{3}{4}$ in. in diameter was suspended by a loop of wire from an arm of a balance, care being

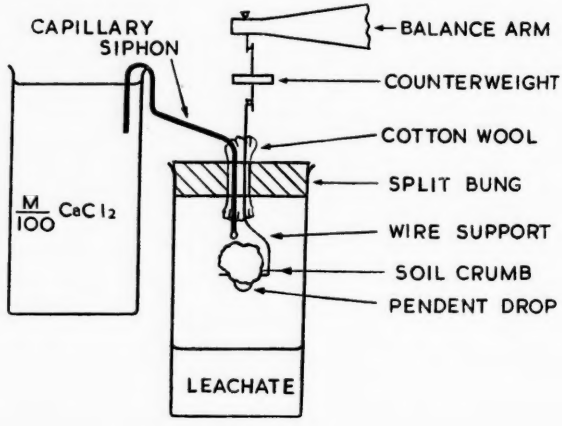


FIG. 1.

taken that the lowest point of the crumb projected below the level of the wire support. A solution of 1×10^{-2} M .$CaCl_2$ was dripped slowly on to the crumb by means of a fine capillary siphon. The height of the outlet of the siphon above the crumb was adjusted so that the pendent drop on the siphon just made contact with the top of the crumb before it was detached. One or two crystals of mercuric chloride were added to the calcium-chloride solution to help to prevent bacterial growth in the crumb.

Experimental Procedure

The air around the crumb was kept as nearly as possible saturated by collecting the leachate in the beaker under the crumb and also by inserting cotton wool in the gaps between the siphon, the hanger, and the cork. The crumb was weighed at the beginning of the experiment and then at intervals until the first drop was detached from the base of the crumb. To weigh the crumb, the cotton wool was removed and the capillary siphon raised. The crumb was weighed daily for the first few days and then at greater intervals of time according to the rate at which the weight was increasing. In order to compare successive weighings, weighings were always made immediately after a drop had detached itself from the base of the crumb. The temperature of the solution in the reservoir beaker was measured after each weighing. Solution was added periodically to the reservoir to maintain an approximately constant head. The rate at which solution was being applied by the capillary tube was determined at the end of the experiment by weighing the amount delivered in 24 hours.

Attempts to determine the volume weight of the crumbs at the end of the experiment were unsuccessful owing to the extreme fragility of the crumbs. One of the grassland crumbs was left to become air-dry, and the volume of air and water plus solids was determined by the paraffin-immersion method (as modified by Puri, 1949). Only the final oven-dry weight was determined on the others. An average figure for the moisture content of crumbs at the wilting-point was obtained by oven-drying large samples. A figure for the subsoil core was obtained by oven-drying two other similar cores.

Results

The relative amounts of water taken up with time by a grassland crumb, a Barnfield 2 N crumb, and a Barnfield subsoil core are shown in Fig. 2. The vertical arrows indicate the times at which the first drop was detached from the crumbs. Details of the crumbs used are given in Table 1.

The residual volume of water referred to in column 6 is the difference between the volume of water taken up after the first drop was detached and the final amount of water taken up. The oven-dry weights in column 3 were calculated, assuming the moisture content at the wilting-point of the particular crumb used was equal to the average moisture content of a large sample of similar crumbs, namely, on an oven-dry basis:

Moisture content of grassland crumbs at pF 4·2 . . . . . = 8·54
,, ,, Barnfield 2 N crumbs at pF 4·2 . . . . . = 9·43
,, ,, subsoil cores (determined on two other cores) . . = 27·3

The volumes of air and solids plus water determined on the grassland crumb after air-drying were 2·44 c.c. and 3·98 c.c. respectively.

From a large sample, moisture content = 4·1 per cent. on oven-dry weight
∴ volume of solids approx. = 9·15 c.c.
and density of crumbs = 2·55.

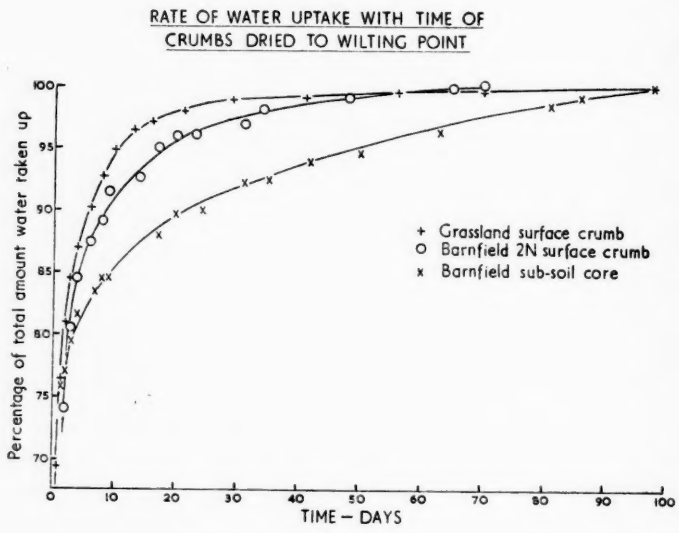


FIG. 2.

TABLE I

Moisture Relationships of Crumbs Used

Description of crumb	*Initial weight of crumb at* pF 4·2 (g.)	*Initial oven-dry weight of crumb* (g.)	*Volume of water taken up when first drop detached* (c.c.)	*Final volume of water taken up* (c.c.)	*Residual volume of water taken up as percentage oven-dry weight of crumb*	*Final oven-dry weight of crumb* (g.)
Grassland .	10·115	9·33	1·863	2·685	9·0	9·15
Barnfield 2 N .	14·529	13·27	1·428	1·933	4·0	12·75
Barnfield 8 O subsoil core.	32·186	25·1	4·624	6·097	6·0	24·7

The rates of application of solution for the three experiments both in c.c./hr. and in c.c./cm.²/hr. are given below. The latter figures for the two crumbs were obtained by treating the crumbs as cubes whose volume was equal to the final volume of oven-dry solids plus water.

Crumb	*c.c./hr.*	*Cross-sectional area: cm.²*	*c.c./cm.²/hr.*
Grassland .	0·080	3·6	0·02
Barnfield 2 N .	0·035	4·0	0·01
Subsoil core .	0·16	18·0	0·01

Discussion

The striking result shown in Fig. 2 is the increase in the weight of the crumbs over a period of 2 or 3 months, even though the crumbs were apparently completely wetted when the first drop formed on the base of the crumbs. The weight of the crumbs was still increasing when the experiment was stopped, although at a much reduced rate. The decrease with time of further water uptake is slower for the subsoil core than for the crumbs. This might be expected from the higher clay content and also the absence of 'cavities', due to organic matter, into which the clay crystals could expand. The amount of 'delayed' uptake, expressed on the oven-dry weight of the crumbs, is twice as high for the grassland as for the Barnfield 2 N crumbs. This could be due to clay crystals in the grassland crumbs being initially in a closer state of packing because of the greater extremes of wetting and drying under permanent grass compared with mangolds, especially as the surface layer of Barnfield tends to waterlog in the winter.

After air-drying the volume of air in the wetted grassland crumbs was 2·44 c.c. If, on wetting, the water just displaced the air and no swelling occurred, 2·44 c.c. of water would be taken up. This is very near the amount of water in the crumb after the first drop was detached, namely 2·54 c.c. It might be inferred, therefore, that up to this time little swelling had taken place.

The practical importance of this delayed wetting lies in its effect on the permeability of the crumbs, and hence on the unsaturated permeability of the soil of which they form part. The rearrangement of the clay crystals as the water films become thicker will lead to a gradual narrowing of the larger pores. Because water conduction depends on the fourth power of the diameter of conducting channels, a small reduction in radius can markedly restrict flow. In the field it might be expected, even after the drains have started to run in the autumn and the soil is apparently saturated, that there should be a slow increase in moisture content down the profile during the winter. This should be accompanied by a slow decrease in the permeability of the profile. Evidence for these two effects, obtained by other workers, will be considered in the following section.

II. FIELD APPLICATION

Increase in Moisture Content after Running of Drains

J. L. Russell carried out a series of determinations of moisture content by weight down to the depths of the drains in Youngman's Pasture (a permanent grass field on the Gault of the Cambridge University Farm) in November–December 1931. He records that the moles ran on 6, 9, and 20 November, when three of the first four profiles were taken. His results are given in Table 2.

As the profiles were taken from different sites, although near each other, comparison of the moisture profiles is complicated by the natural heterogeneity of the soil. However, ignoring the readings for the surface

TABLE 2

Moisture Profiles of Youngman's Pasture, 1931

	Water content—percentage weight of dry soil					
Sample number	*I*	*II*	*III*	*IV*	*V*	*VI*
Date of sampling	6 Nov.	9 Nov.	13 Nov.	20 Nov.	28 Nov.	12 Dec.
Depth of sample (in.)						
0–1	61·0	86·0	58·3	44·8	77·5	69·1
1–2	45·2	53·7	52·8	50·5	66·1	61·2
2–3	40·0	47·1	46·0	43·7	56·0	55·6
3–4	35·7	41·8	41·3	39·5	43·6	47·2
4–5	33·4	32·0	39·0	35·7	41·5	42·2
5–6	32·0	33·8	39·0	34·0	37·0	40·9
6–8	36·1	33·3	35·8	31·8	36·0	38·7
8–10	31·1	25·1	35·5	32·1	33·8	36·3
10–12	30·8	29·4	29·8	32·2	30·8	33·7
12–16	29·6	29·2	31·3	30·9	32·9	31·9
16–20	29·8	29·7	32·3	33·3	33·6	34·2
20–24	31·4	32·3	32·8	33·9	36·2	35·0

The rainfall over this period, in inches, was as follows:

November .	3	4	6	7	9	10	11	
Rainfall . .	0·02	0·09	0·40	0·04	0·21	0·12	0·12	
Sample no. .			I (before rain)		II (during and after rain)			
November .	13	18	20	23	25	26	27	28
Rainfall . .	..	0·49	..	0·11	0·04	0·15	0·30	..
Sample no. .	III		IV					V
December .	2	3	5	9	12			
Rainfall . .	0·04	0·09	0·36	0·03	..			
Sample no. .					VI			

4 in., where oven-dry moisture contents will be particularly variable owing to the large mass of roots present, the figures in Table 2 have been analysed statistically, yielding the following results:

Means Over All Depths versus Times

Profile	I	II	III	IV	V	VI
Mean	31·8	31·1	33·8	33·0	35·0	36·8

With a standard error of 0·58, a difference between the means of 1·74 will be significant at the 5 per cent. level. There is every reason to believe, therefore, that there is a genuine increase in moisture content with time at all depths, although the soil was sufficiently wet for the moles to be running when the first samples were taken.

Drainage Records

One method of determining the variation in permeability of a clay soil over the winter is by following the change in shape of the rate of flow-time curves of the drains. Two idealized drainage curves are shown in Fig. 3.

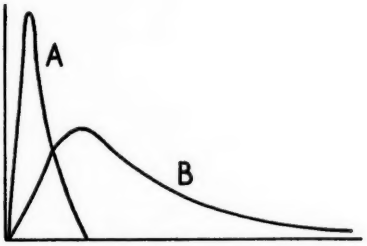


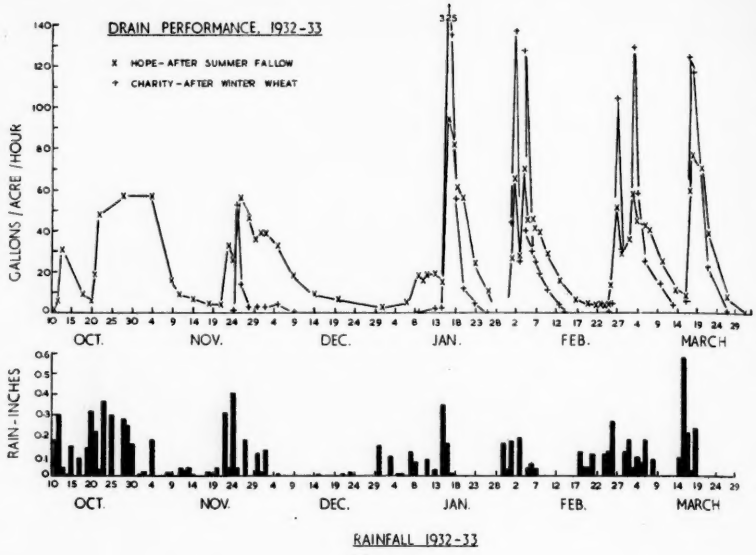
FIG. 3. A = soil with high permeability.
B = soil with low permeability.

For the same total amount of water discharged, the rate of flow from a soil with a high permeability has a very high initial peak and then drops rapidly to zero (curve *A*), while with a soil of low permeability the initial peak is much lower and slow drainage occurs for several days.

H. H. Nicholson measured instantaneous rates of flow from the tiled outlets of several mole-drained fields on the Cambridge University Farm during the winters of 1932–3 and 1935–6. His results for three fields are plotted in Figs. 4, 5, and 6, together with rainfall figures over the two periods. The fields are all on the Gault, which is naturally impermeable below the depth of the drains, so that no through percolation takes place. The cropping of the fields was as follows:

Field	*1931*	*1932*	*1935*
Hope. .	Mole-drained and fallowed	Fallowed. Silage drilled in Nov. 1932	. .
Charity .	Mole-drained and fallowed Winter wheat sown Nov. 1931	Wheat harvested Aug. 1932 Winter oats sown Oct. 1932	Winter wheat Oct. 1934
Youngman's Pasture	Old permanent pasture mole-drained Sept. 1931.		

The results for the winter of 1931–2 afford a very good test of the effect of drying by roots during the summer on soil permeability since Charity and Youngman's Pasture both carried a crop during the summer while Hope was fallowed. Referring to Fig. 4, where the drainage curves for Hope and Charity are plotted, the drains on Hope ran considerably earlier than those on Charity, reflecting the much greater loss of water by evaporation from the cropped soil compared with the uncropped soil during the summer. Once the drains on Hope had started to run they continued to do so throughout the winter. On the other hand, those on



RAINFALL 1932-33

FIG. 4.

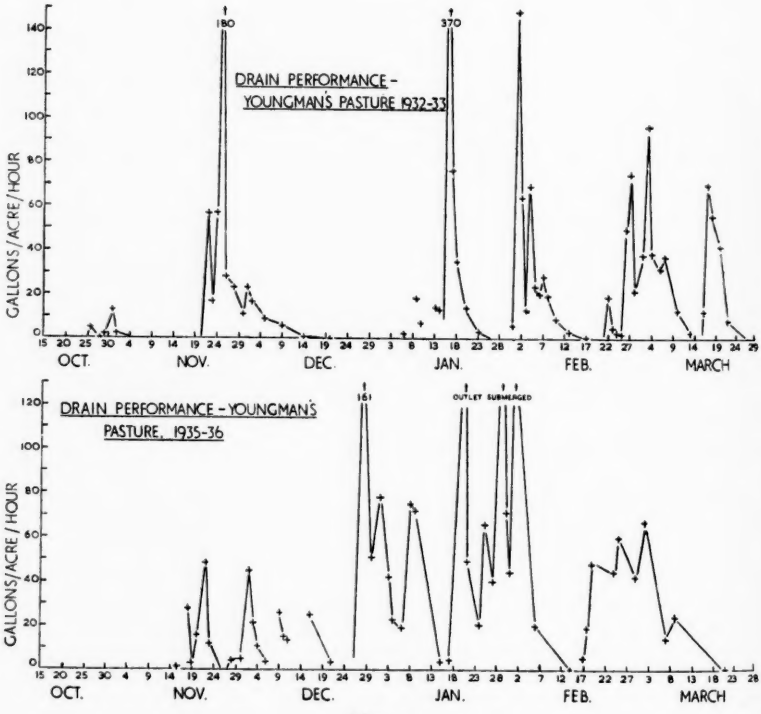


FIG. 5.

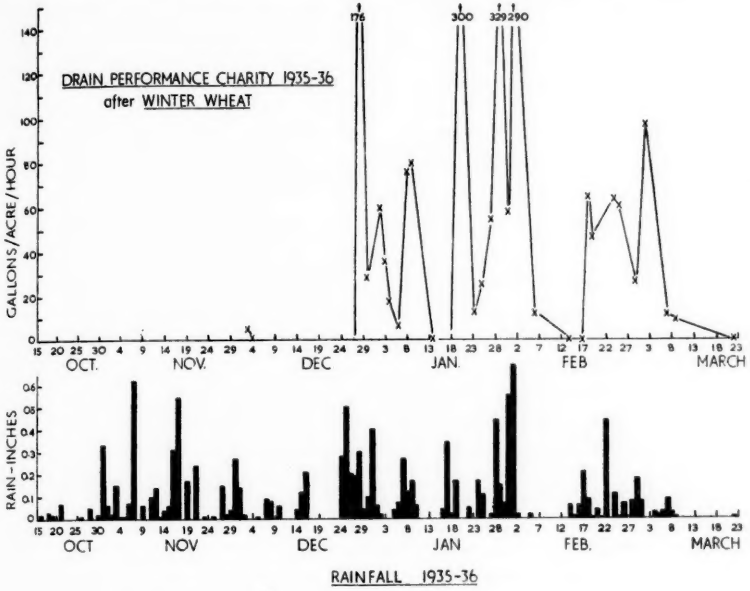


FIG. 6.

Charity gave very rapid flushes followed by periods of no flow. The difference is particularly marked in the flow patterns after the two periods of heavy rain, 15–16 January and 30 January to 3 February. In fact the two sets of curves correspond very well with the 'ideal' curves *A* and *B* in Fig. 3 for soils with high and low permeability respectively. By the middle of March the difference between the two drainage curves had become much less marked. Although the peak rate of flow from Charity was still higher than that from Hope, the drains continued to run for nearly the same length of time, indicating that most of the delayed swelling had taken place.

The volume of water drained in a given time is equal to the area under the rate of drainage-time curve over this time period. It has been shown by Penman (1948) that evaporation from a bare soil during the winter, provided the surface is well wetted, depends only on meteorological factors. As drainage is the excess of rainfall over evaporation, it might be expected that equal volumes of water would be discharged from Hope and Charity once both sets of moles were running. The moles on Faith will, of course, begin to run considerably earlier than those on Charity owing to the much greater loss of water by evaporation during the summer from a growing crop than from a bare fallow. However, by measuring the areas under the curves, the difference in the volume of water drained from Hope and Charity in the 50 days period 24 November–14 January, immediately after the moles in Charity had started to run, has been found to be $1{\cdot}9 \times 10^4$ galls./acre.

To show that this amount of water can be reasonably accounted for by swelling, upper and lower values for delayed water uptake have been worked out on the basis of laboratory results for the Barnfield subsoil core; the former based on the water uptake over the *first* 50 days after the first drop had been detached from the base of the core (see Fig. 2), the latter on that over the *last* 50 days (i.e. 50–100 days). The results for the subsoil core were preferred to those for surface crumbs, as being more representative of the soil profile as a whole.

Amount of further water uptake in first 50 days = 4·75 per cent. of oven-dry weight
" " " " last " = 1·25 " "
Weight of an acre of soil down to drains (at 24 in.) = 4,000 tons
Upper value to delayed water uptake per acre = $4{\cdot}3 \times 10^4$ galls.
Lower " " " " = $1{\cdot}1 \times 10^4$ "
(These values are equivalent to 1·9 in. and 0·5 in. of rain respectively.)

The first case would correspond to a series of very heavy falls of rain at the end of the summer, so that drainage occurs soon after the upper layers of the soil are wetted. The second case will be approached when the soil is slowly wetted in the autumn by successive showers of rain and there has been time for most of the soil to swell appreciably before the 24-in. layer is sufficiently moistened for the moles to run. It will be seen that the figure given in the previous paragraph lies satisfactorily within the above limits.

At the end of the winter, when swelling would be expected to be practically complete, the amount of drainage from the two fields should be nearly equal. From the areas under the graphs, the volumes of water drained over the period 15–31 March have been found to be $1{\cdot}2 \times 10^3$ and $1{\cdot}1 \times 10^3$ galls. per acre from Hope and Charity respectively. Although this difference is small, taken in conjunction with the slight difference in the shapes of the two crumbs already mentioned, it does indicate that even at the end of this particular winter the clay in Charity was not quite fully swollen.

The drainage record of the moles in Youngman's Pasture for the same winter, 1932–3, is shown in the upper half of Fig. 5. The curve is incomplete in places owing to lack of data. The flow patterns are similar to those for Charity, namely rapid flushes followed by sharp cut-offs, indicating a highly permeable soil. Curves for Youngman's Pasture and Charity (again after winter wheat) for the winter 1935–6, together with the rainfall for this period, are shown in the lower half of Fig. 5 and Fig. 6. Again, as in 1932–3, the curves are similar, being characterized by high initial rates of flow and sharp cut-offs, the heights of the initial flushes and the sharpness of the cut-offs decreasing as the winter progresses. It will be seen that in both years the initial flushes are higher and subsequent cessation of drainage occurs sooner on Charity compared with Youngman's Pasture. Also the drains on the grassland started to run earlier, and once both were running a greater volume was discharged from the grassland. In the light of the previous paragraph, this would indicate that the clay swells more quickly under grass than after wheat, or that the wheat roots dry out the subsoil more efficiently. Some evidence for this was found by Teixeira (1946), who dug profile pits in adjoining

fields at Rothamsted, one in permanent pasture and the other on a wheat stubble. His qualitative estimation of the relative abundance of roots with depth is as follows:

Grass		*Winter Wheat*	
Depth (cm.)		*Depth (cm.)*	
0–15	Great numbers	0–20	Great numbers
15–30	Many	20–45	Great numbers
30–50	Roots	45–105	Roots
70	Only fine	105–135	Roots
80	None	135–160	Few roots

His results clearly show the greater abundance of wheat roots in the lower parts of the profile.

It would appear that one method of estimating delayed swelling and its dependence on cropping would be to examine the drainage records of a series of parallel plots.

Field Capacity

Field capacity is defined as the amount of water held in the soil after rapid drainage has ceased. In practice this is taken to mean 2 or 3 days after rain. On account of delayed swelling, field capacity in the autumn will correspond to a lower soil-moisture content than field capacity at the end of the winter. It is interesting to note that Kraebel and Sinclair (1940), investigating the dependence of field capacity on depth of wetting, found that the surface soil only reached its maximum moisture content when it had been wetted to at least a depth of 30 in., i.e. the clay in the surface soil had been given time to swell.

In recent years Penman has used the deficit of rainfall below evaporation, starting from a date in the spring at which the soil is at field capacity, to calculate when irrigation is required. No ambiguity arises in this case because, as has already been noted, swelling is practically complete by the end of March. However, Penman (1949) attempted to predict the date in the autumn for the years 1933–40 on which the mole drains of Youngman's Pasture should start to run. He assumed that this would occur when the deficit was again reduced to zero, i.e. when the moisture content of the soil in the autumn was equal to that of the preceding spring. His apparent success was due to the adjustment of the root constant, a semi-empirical constant which is used in the computation of evaporation from meteorological data to make the data fit. From the approximate calculation given earlier it is evident that it may be only necessary to adjust the constant by 0·2 in. to mask the effect of swelling, whereas the values used varied from 2·5 to 3·6 in. in different years. However, there were occasions in which the predictions were completely unsuccessful. In 1939–40 a root constant of 1·7 in. would have had to be taken to forecast the correct date, that is about 1·5 in. below the mean value of the root constant. In that year the drains ran earlier than in any other year, giving the least time for the clay to swell. As has been seen already, the difference of 1·5 in. could easily be taken up by swelling.

The other occasion was the summer running in 1936 when unusually heavy rainfall was sufficient to cause the drains to run; and as there was little time for the clay to swell there would be an appreciable deficit based on spring field capacity. (Penman calculated 0·7 in.) It is quite possible that if swelling is taken into account the root constant may be found to be approximately constant over all the years. This would simplify calculation of actual evaporation.

Penman and Schofield (1941) compared the performance of the Rothamsted drain gauges. They reported that, after dry summers, the volume of drainage during the following winter from the 20-in.-deep gauge was persistently greater than that from the 60-in.-deep gauge. From the records, the maximum differences in the volume of water drained once both gauges were in action in the period 1880–1930 have been found to be 0·84 in., 0·85 in., and 0·87 in. in 1911, 1914, and 1929 respectively. Also, in these particular years the rainfall in September and October was very low, so the clay had little chance to swell before appreciable percolation occurred. The difference in drainage volume is ascribed to slow water uptake by the clay below a depth of 20 in. in the 60-in. gauge. Using the value of 6 per cent. by weight (see Table 1) for delayed swelling found for the Barnfield subsoil core (the drain gauges are isolated blocks of Barnfield soil) and the value given by Lawes and Gilbert (1871) for the bulk density, 0·87 in. of water would be taken up by 40 cm. depth of clay. This corresponds to a total depth of drying in the 60-in. gauge of 91 cm.

This is in good agreement with the results of Keen (1939), who followed the retreat of the water table in a 6-ft. tank filled with similar soil when exposed to natural evaporation. During a prolonged drought Keen found that after 2 months the water table had fallen to 80 cm., but after a further 4 months' drying the water table was almost stationary at about 93 cm. below the soil surface.

Crumb Cohesion

As already mentioned at the beginning of the paper, the swelling of crumbs will also be accompanied by a decrease in cohesion, the cohesion becoming vanishingly small as the suction approaches zero (unless there are inorganic cements, e.g. iron oxide, or organic cements present). The persistence of mole channels in clay soils may be due to the fact that the crumbs forming the walls never swell up completely. In this connexion it is known that moles are much more likely to function satisfactorily if pulled in the spring rather than the autumn. This is probably because the spring mole channels become thoroughly dried out in the following summer, through the opening of the mole slits as the clay shrinks, before the moles begin to conduct water in the autumn.

The reduction in cohesion with swelling may also be important in some engineering projects. Railway cuttings through the London clay in Essex frequently slide after a few years (Toms, 1948). This could be due to the slow swelling of the initially dry clay after the overburden has been removed.

Summary

Soil crumbs dried initially to the wilting-point have been slowly percolated with 1×10^{-2} M. calcium chloride using a capillary siphon. It has been shown, by weighing the crumbs at intervals, that although excess solution is discharged from the base of the crumbs after the first or second day, the crumbs continue to take up solution at a decreasing rate over the next 3 months. This is attributed to the rearrangement of the individual clay crystals as the thickness of the water films on the crystals increases with decreasing suction.

The slow swelling of the clay will be accompanied by a corresponding decrease in the permeability of the crumbs as the larger pores are reduced in size. This is of great practical importance in the drainage of heavy clay soils, since it means that the permeability of the soil as a whole may be improved over the winter by using a deep-rooted crop to dry the subsoil out during the summer. This has been shown by comparing the drainage records of three fields; one after fallow, one after winter wheat, and the third in permanent grass on the Gault of the Cambridge University Farm for the winter of 1932–3. It was also possible to infer from the dates at which the drains started to run and the volume of water discharged that considerable delayed water uptake occurs as the clay slowly swells.

Delayed swelling also affects the concept of field capacity, since the moisture content of the soil at 'field capacity' will be appreciably higher after a long wet period than when through percolation first occurs. Penman's (1948) attempt to forecast the running of mole drains has been re-examined in this light.

The time taken for clay to swell partly explains why mole draining is successful at all and why it is preferable to mole in the spring rather than in the autumn. The gradual reduction in the cohesion of crumbs as the clay slowly swells may be important in some engineering problems.

Acknowledgements

I am greatly indebted to Mr. H. H. Nicholson, of the School of Agriculture, Cambridge, for permission to use his drain-gauge results, and also to Mr. J. Dunwoody for his statistical analysis of the results given on page 152 of this paper.

REFERENCES

EMERSON, W. W. 1954. In the press.
KEEN, B. A. 1939. Q. J. Roy. Met. Soc. **65**, 123.
KRAEBEL, C. J., and SINCLAIR, J. D. 1940. Trans. Amer. Geophys. Union, **21**, 84.
LAWES, J. B., and GILBERT, J. H. 1871. J. Roy. Agric. Soc. **7**, 91.
NICHOLSON, H. H. Unpublished data.
PENMAN, H. L., and SCHOFIELD, R. K. 1941. J. Agric. Sci. **31**, 74.
—— 1948. Proc. Roy. Soc., A., **193**, 120.
—— 1949. J. Soil Sci. **1**, 74.
PURI, M. L. 1949. Ph.D. Thesis of London University.
SCHOFIELD, R. K. 1946. Trans. Faraday Soc. **42 B**, 219.
TEIXEIRA, A. J. DA S. 1946. Ph.D. Thesis of London University.
TOMS, A. H. 1948. Proc. Conf. on Biology and Civil Engineering, p. 99. (Institution of Civil Engineers.)

A NOTE ON THE SODIUM SATURATION TEST FOR DETERMINING THE COHESION OF MOIST SOIL CRUMBS

W. W. EMERSON

(*Rothamsted Experimental Station*)

A NEW method of determining the cohesion of moist soil crumbs has recently been described by Emerson (1954*a*, *b*). The method consists of replacing all other exchangeable ions on 20 g. of 'natural' air-dry 1–2 mm. soil crumbs with sodium, and then measuring the decrease in the permeability of the crumb bed when brought into equilibrium with successively more dilute solutions of sodium chloride. In the light of further work two modifications to the technique have been found necessary. The apparatus, however, remains basically as drawn in the previous paper in this Journal (1954*a*).

Porosity of the funnel

The decrease in the permeability of samples from the same field with decreasing salt concentration was found to vary markedly with the particular sintered glass funnel used. This was attributed to the clogging of the finer pores in the sinter by clay and silt particles dispersed from the soil crumbs. Substituting coarser 17 × 1 funnels (as supplied by Messrs. Gallenkamp, London), the sinters of which have a minimum pore size of about 20μ, has removed this difficulty. At the same time these funnels will withstand about 19 cm. of water suction, so that the initial slow wetting of the crumbs at 12·5 cm. suction, to eliminate breakdown due to slaking, may still be carried out. From data supplied by Messrs. Gallenkamp the variation in permeability of the funnels themselves is about 5 per cent.

Constancy of hydraulic head

It has been found that at the lower salt concentrations where the permeability of the crumb bed has decreased due to swelling and dispersion of the crumbs, lowering the outlet and therefore increasing the hydraulic head actually lead to a reduced rate of flow. This is, of course, contrary to experience with a bed of rigid particles, but with weak soil crumbs the crumbs may be squashed by the increased suction applied to them as the hydraulic head is increased.

In fact, the total pressure in the crumbs is the sum of the repulsive pressure between the clay crystals in the dilute salt solution and the hydraulic pressure which is equal to the difference in height between the crumbs and the outlet. The contribution due to the hydraulic head becomes more important as the salt concentration is reduced. For example, percolating the crumb bed with distilled water, the equilibrium concentration of sodium chloride in the leachate is about 0·3 mN. From known data on Wyoming bentonite and using the usual electric double-layer theory, the repulsive pressure between the clay crystals at this salt

concentration is about 2 cm. of water. This also suggests that the relative strengths of crumbs which have not completely dispersed after percolating them with distilled water could be further differentiated by measurements of permeability at increasing heads.

In practice a head of 1 cm. has been found convenient and this is maintained constant throughout. Using the 17×1 funnels, this leads to very high rates of flow before the crumbs begin to break up. It is necessary, therefore, after measuring the initial flow rate, to adjust the tap to maintain a flow of about 9 c.c. in 45 sec. Next morning, of course, when the permeability is again measured, the tap is kept fully open.

Sodium saturation

The rate at which calcium and magnesium ions are displaced from 20 g. of 1–2 mm. soil crumbs by 500 mN. NaCl has been investigated for two Rothamsted soils, Sawyers 1 and Barnfield plot 8 O. The former is an acid soil, the latter contains free $CaCO_3$. 500 mN. NaCl was slowly percolated through the crumb bed at a rate of 500 c.c. in 24 hours, and the concentration of divalent cations in the leachates determined by titration with versenate. The following table gives the molar concentrations found in successive 500 c.c. of the leachates:

	Day 1	*Day 2*	*Day 3*
Sawyers 1 . . .	$2{\cdot}60.10^{-3}$	$3{\cdot}0.10^{-3}$	$7{\cdot}0.10^{-5}$
Barnfield 8 O . . .	$3{\cdot}45.10^{-3}$	$1{\cdot}06.10^{-3}$	$4{\cdot}5.10^{-4}$

The free calcium carbonate in the Barnfield soil would maintain a concentration of about $5{\cdot}10^{-4}$ m. Ca in equilibrium with atmospheric CO_2. It will be seen that the 2-day leaching used previously is adequate to replace the exchangeable cations in the soil with sodium.

The series of salt solutions 100 mN., 20 mN., and 2 mN. NaCl followed by distilled water (1954*b*) for assessing the cohesion of crumbs has been found to be satisfactory. With the coarser funnels and keeping a constant hydraulic head, reproducible values of the permeability at these salt concentrations have been obtained on samples of the same soil. The test is now considered suitable for routine measurements.

Thanks are due to Mr. F. M. Axon for carrying out the versenate titrations.

REFERENCES

EMERSON, W. W. 1954*a*. The determination of the stability of soil crumbs. J. Soil Sci. **5**, 233.

—— 1954*b*. The effect of various grasses and lucerne on the cohesion of soil crumbs. Trans. Fifth Internat. Congr. Soil. Sci. (In the press.)